State Univ. of N.Y

Global Tectonic Studies

NSG-5274

Final Technical Report

Hotspots and Anomalous Topography

(1) Hotspot studies A comprehensive catalogue of hotspots with maps and detailed descriptions has been prepared. A considerable emphasis has been on the relationship of hotspots to the more extensive flood basalts and papers on this topic were presented at the Hawaiian symposium (Encl. 1) and contributed to the Wilson celebration (Encl. 2 preprint).

(2) Anomalous topography Analyses of the topography of the continents are being compiled and an example dealing with South America (Encl. 3) is provided. Studies of plateaus were reported at the plateau conference in Flagstaff (Encl. 4) and problems of renewed uplift of old mountain belts at the US-Yugoslav conference on intracontinental earthquakes (Encl. 5).

Kevin Burke

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Enclosure 1

Abstract Volume

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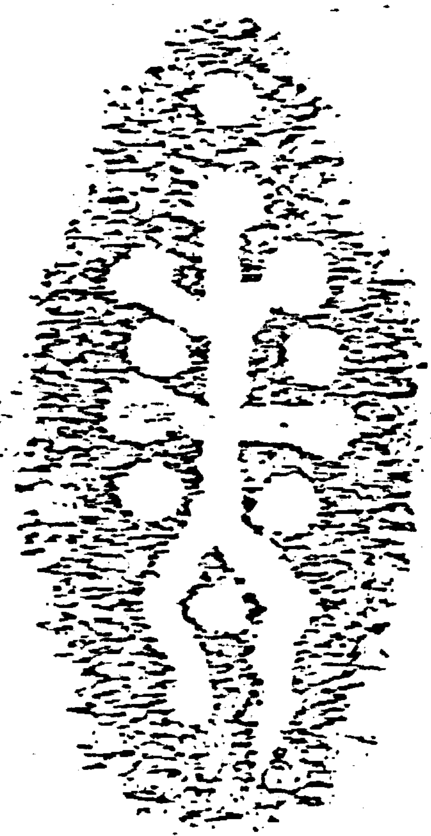
Hawaii Symposium on Intraplate Volcanism and Submarine Volcanism



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HILO, HAWAII
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THE DISTRIBUTION OF HOT SPOTS AND HYPOTHESES FOR THEIR ORIGIN

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Most of the vulcanism on earth occurs along plate boundaries; in oceanic spreading ridges, in island arc/cordilleran (Andean) zones, and in Tibetan plateau-type collisional zones. Basic rules governing the origin and distribution of these three types of volcanic activity are reasonably well defined. Other vulcanism, collectively called hotspot vulcanism, is commonly attributed either to cracking of the lithosphere and consequent rise of magma produced by partial melting of the mantle rising to fill the cracks (passive mantle hypothesis), or to chemical or physical inhomogeneities in the mantle (active mantle hypothesis). We have compiled a global map and catalogue of hot spots to help in testing the two classes of hypotheses. Although there is a great range in volume of vulcanism, amount and diameter of associated uplift, and duration of activity, there is little evidence that more than one population of objects is present. The crucial fact about the hot spot distribution is that a number occur right on the axes of spreading ridges (e.g. Iceland, Galapagos, Azores). These retain their contrast with adjacent "normal" oceanic spreading ridge (distinctive chemical and isotopic compositions, anomalous elevation, and excess volume of magmatism) despite the negligible lithosphere thickness. Some have remained approximately axial, shown by symmetrical traces, for up to 100 m.y., and they appear to control the ridge position (ridge jumping, e.g. Iceland, Galapagos). These facts are strong evidence against the various crack propagation hypotheses and the anchor-asperity model for hot spots. A thermal and /or chemical disturbance in the mantle, as originally suggested by Wilson, seems the only general hypothesis that will account for most hot spots. A few have been proposed to be the result of tensional zones generated by continental collision, but for these cases opportunism (formation of tensional zones in areas of pre-existing thinned lithosphere) has not been ruled out. A single hypothesis for hot spots (active mantle) seems inherently preferable to the many passive mantle hypotheses that have been proposed.

HOT SPOTS, HIGH SPOTS, AND OTHER TYPES OF NON-PLATE MARGIN (ANOMALOUS) TOPOGRAPHY

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A global compilation of hot spot vulcanism reveals that most active hot spots are associated with topographically high and structural elevated areas. The continental portion of the African plate displays this relationship particularly clearly, since there are a large number of hot spots on it and they are relatively well mapped. Burke and Wilson proposed that Africa has moved little over the past 25 m.y. with respect to any of the slowly moving reference frames, and we suggest that this has enabled the effects of mantle disturbances to be better displayed on this plate compared with most other areas of the earth. The diameters and amplitudes of the uplifts, and the amount of vulcanism associated with African hot spots range from large values to small, but they are not obviously correlated, since large and/or high structural uplifts are not necessarily capped by much, or any, volcanic accumulation. Thus, although the vulcanism is perhaps the most prominent feature of the hot spots, we suggest that it is of secondary importance compared to the structural uplifts (high spots) and the major active intracontinental sedimentary basins that some of them surround. Burke and Whiteman suggested that hot spot uplifts were all around 200 km across; some (e.g. Ahaggar) are somewhat larger (500 km), although a majority of the African population are close to 200 km in average diameter. Larger swells around 1000 km across seem to us to consist of groups of hot spots, although the East African swell has a lesser but still anomalous elevation between those hot spot uplifts found within it. Most of the uplift-defined hot and high spots in Africa are considerably smaller than this and cannot be readily grouped into such large 1000 km diameter swells; the same is true for a majority of the global hot spot population. A more realistic description allows a wide range of diameter and amplitude for hot spot type uplifts. Examples of high spots without associated vulcanism exist on other plates; prominent continental examples are the Adirondacks and the Putorana massif. Sub-oceanic examples include the Bermuda Rise and the northeast Pacific uplifts found by Menard. However, they (and hot spots) are almost everywhere much less prominent than on the African plate, which, we suggest, is at least partly a function of plate velocity with respect to the local mantle. High spots, like hot spots, are difficult to detect in zones of convergent tectonics and this includes the anomalous topography of Central Asia, proposed to be the result of the Indian-Asian continental collision. Other types of intraplate anomalous topography are elongate and readily distinguished and they include old collisional orogenic belts up to about 0.5 b.y. old (e.g. Urals, Appalachians), and forebulges due to lithospheric flexure in front of active thrusting zones (e.g. central India). Other unusual elongate zones of anomalous topography include one due to recent incipient convergent tectonics (S. India-Ceylon) and one (E. Baffin Island-N. Labrador) for which there is no obvious explanation or counterpart. Lithospheric thinning is the most plausible explanation for the uplifts associated with hot spots, a process that seems readily driven by mantle upwelling under the hot and high spot sites.

African Hotspots and Their Relationship to the Underlying Mantle

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Several lines of evidence indicate that Africa may be at rest with respect to the underlying mantle, and has been for the past 25 million years. This would allow weaker mantle plumes to pierce the overlying plate, creating recognizable hotspots or areas of uplifts without volcanics (Highspots). Also, the surface expression of the hotspots will not be blurred by large motions relative to the underlying mantle. Therefore, the African Plate may provide a more complete picture of the underlying mantle convection pattern then plates in motion would.

I have analyzed the hotspot population of Burke and Kidd (1975) for the African Plate and also the highspot distribution (Kidd, personal communication), using the polygon method of Thiessen (1911). A polygonal convection pattern is indicated which in planform closely resembles laboratory convection models under a stationary plate. When the nearest neighbor separation distances are scaled to simulate mantle convection extending the the 700 kilometer seismic discontinuity, the lab models show the same separations as the hotspots and the highspots do. It is concluded that this type of convection system is causing the hotspots and highspots of Africa, and possibly elsewhere in the world.

Enclosure 2 5

VOLCANISM ON EARTH THROUGH TIME

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ABSTRACT

Volcanism is widespread on Earth and apparently always has been. In this respect the Earth contrasts with the Moon, where volcanism stopped about 3 Ga ago, and with Mars, where it may have stopped 1 Ga ago. Although the terrestrial volcanism that occurred before the oldest preserved rocks formed was probably similar to later volcanism, the very earliest volcanic activity on Earth could have been like that on the Moon, if the Earth acquired its water late during the high impact flux.

Volcanism plays a vital part in all three stages of lithospheric evolution active on Earth today. Basalt forms at divergent plate boundaries; tholeiitic and calc-alkaline rocks, most characteristically andesite, dominate where arcs form above subduction zones; highly potassic volcanics are associated with active continental collision, crustal thickening and fractionation. These three processes appear to have operated at plate margins throughout most of the Earth's history.

Ultramafic komatiites, forming perhaps 5% of basalt piles and indicating the existence of ultramafic melts, are peculiar to the Archean, and are presumably due to the greater heat generation of the early Earth, although their precise significance is ambiguous. Non-plate margin (hot-spot) alkaline and tholeiitic volcanism is also recorded in old rocks. Remnants of flood basalts and associated sills and dike swarms formed by rifting episodes are clearly displayed in the Canadian shield up to 2.5 Ga ago, and in Greenland back to 3.6 Ga.

Sea level variations during the Phanerozoic have been interpreted as indicating variations in spreading ridge volume and hence (on an Earth of roughly constant volume) of plate creation and destruction rates. Precambrian sea level fluctuations cannot be interpreted in the same way because the sediment record is incomplete and because timing has not been adequately resolved. Estimates of episodicity in plate activity in the Precambrian depend mainly on the occurrence of peaks in age abundance. This method is highly misleading because the preserved areas of particular ages are much too small to be representative of the world at those times. As far back as the geological record goes, volcanism has been continuously active on Earth and has occurred mainly at plate boundaries. Phanerozoic history indicates some episodicity, and the Precambrian was probably similar.

INTRODUCTION: VOLCANISM ON EARTH COMPARED WITH THAT ON OTHER TERRESTRIAL PLANETS

Active volcanism is widespread on Earth and is recorded in rocks of all ages. Strikingly, the oldest preserved rocks on Earth (at Isua in West Greenland) appear to be volcanic breccias. With the exception of a very small volume produced by partial melting following impact (e.g. basaltic rocks preserved in plutonic facies at Sudbury and the Bushveld, and suevites at numerous localities) terrestrial volcanic rocks appear to be mainly the products of partial melting generated fundamentally by the decay of radioactive heat-generating nuclides.

In this respect the Earth contrasts with some other terrestrial planets and satellites. On the Moon, volcanism ended about 3 Ga ago, and much, if not all, of lunar volcanism was related to impacts ~4 Ga ago or earlier. By 3 Ga ago, radioactive heat generation within the Moon seems to have declined to a rate sufficiently low to have been entirely removed through the surface by conduction. This suggests to us that the Moon's heat-generating nuclides were concentrated close to the surface by that time.

On Mars, earlier volcanism appears to have been impact-dominated. However, the spectacular later volcanism of the Tharsis area and some older similar features more likely resulted from partial melting through radioactive decay of heat-generating nuclides. Martian volcanism seems to have ceased, but there is some uncertainty as to exactly how long ago this happened. We do not yet know about volcanism on Venus, but extremely violent volcanic activity is in progress on Io (Morabito, et al., 1979). This volcanism, perhaps unique in the solar system, appears to be the result of tidal forces (Beale, et al., 1979). Tidal forces do not appear to be major generators of volcanic heat elsewhere in the solar system, although tidal influences triggering eruptions have been recognized on the Earth

(e.g. Mauk and Johnston, 1973).

SECULAR VARIATION OF TERRESTRIAL VOLCANISM

This short review deals only with volcanism, but plutonic igneous rocks are associated with all volcanic rocks, and most plutonic rocks exposed or emplaced at high crustal levels at one time underlay volcanic rocks. In a few, probably exceptional, areas shallow intrusives do not appear to be linked to volcanic rocks. Examples exist in the young high-level granites of the Himalayas (Gansser, 1964), where the absence of volcanic superstructures may be due to intrusion in a dominantly compressional environment, or to rapid and effective erosion following uplift. This contrasts with conditions on the Tibetan plateau to the north, where compression is now less dominant and young volcanism has been widespread.

Estimates of the variation of heat generation in the Earth through radioactive decay with time are shown in Figure 1. Small additions through such processes as tidal dissipation are significant through all of geologic time, and much larger contributions to heat generation were made before 4 Ga by core formation and impact. Like Turcotte and Burke (1978) we assume that heat escapes from the Earth roughly as soon as it is generated, and generally within 0.5 Ga.

Volcanism is, and apparently always has been, one of the dominant processes by which heat is removed from the Earth's interior. At present, heat leaves the Earth in three main ways: by conduction; by the eruption and rapid cooling of igneous rocks, and by the longer-term process of the cooling of ocean floor as it ages, subsides and moves across the Earth's surface. We have shown elsewhere (Burke and Kidd, 1978) that because some of the ancient continental crust is and was of normal thickness and its base did not suffer widespread melting, the continental conductive thermal gradient about 2.5 Ga. ago in the lower crust was not significantly greater than it

is now. Since Figure 1 indicates that two or three times as much heat was being generated at that time, this greater terrestrial heat must have been removed by the other two processes: igneous rock emplacement and ocean-floor aging. Ocean-floor aging is linked to divergent plate boundaries; most other igneous activity today also occurs at plate boundaries.

We and our colleagues (Burke and Dewey, 1972; Dewey and Burke, 1973; Burke, et al., 1976; Burke, et al., 1977) have discussed at length elsewhere why we consider most igneous activity to have been plate-boundary related throughout recorded geologic history (about the last 3.8 Ga). Others have interpreted the early lithosphere as not broken into rigid plates and consequently most tectonic and igneous activity as unrelated to plate margins. It is unnecessary to repeat the arguments for and against these interpretations. Let it suffice to say that the rocks and structures formed in early times so closely resemble later ones, that any radically different process operating then must have been capable of simulating the results of present plate tectonics. It is mainly for this reason that we prefer to regard the ancient world as broken into plates. There are two minor but significant differences between early and present-day plate tectonics. First, the size of suture-bounded areas in Archean terranes (notably the Superior Province) is generally smaller than in later terranes. This is consistent with the idea that greater lengths of plate boundary and/or faster moving plates were needed to dissipate the extra Archean heat (Burke and Kidd, 1978). Secondly, Archean terranes contain most known examples of a unique class of volcanic rocks, the ultramafic komatiites, which form a small proportion of Archean basaltic piles. Archean komatiites are presumably related to the greater heat generation of the early Earth and their restricted occurrence has been used by various workers as a datum point in analyzing the thermal history of the Earth. Clearly, this is qualitatively correct (the Earth was hotter),

but we know too little about the details of the origin, environment and mechanism of eruption of the Archean komatiites to make quantitative inferences about the significance of these peculiar rocks. In summary, we conclude from the composition, distribution and structure of volcanic rocks occurring among the oldest of terrestrial rocks that most volcanic activity from 3.8 Ga on was, as now, related to plate margins.

VOLCANISM AT PLATE MARGINS: DIVERGENT MARGINS

There are three types of plate boundaries (Wilson, 1965) and it is appropriate to consider volcanic activity associated with each of them. Tuzo Wilson has long advocated the integration of solid-earth sciences, pointing out that geology, geophysics and geochemistry are not realistically separable, and are strongest when results from all fields are considered together. There is no better example of this integration than the study of divergent plate-margin volcanism (which takes place at oceanic spreading ridges and, except in anomalous places like Iceland and the Afar, occurs two or more kilometers below sea level). Geophysical methods have dominated the study of oceanic spreading ridge volcanism and have been complemented by geochemical study of dredged pillow-lavas and limited submersible reconnaissance. The realization that ophiolite sequences in mountain belts represent small samples of ocean floor and preserve material produced at divergent plate boundaries has permitted integrated studies of the type favored by Wilson, that have done much to reveal the essential features of divergent plate-boundary magmatism.

Figure 2 shows the results of one study of this kind (based on Dewey and Kidd, 1977). Basalt rising from partially melted mantle occupies a magma-chamber roughly triangular in cross-section; volcanic material is erupted from this chamber to form the pillow lavas of the ocean floor,

the underlying sheeted dikes and the cumulate and non-cumulate plutonic rocks that overlie the depleted mantle.

The recognition of discrete major volcanoes within the axial valley of the FAMOUS area (Ballard and van Andel, 1977; Ramberg and van Andel, 1977) and across-strike compositional variations in basalt (Bryan and Moore, 1977) has led to interpretations of slowly spreading boundaries having along-strike diversity and episodic structural development. This kind of interpretation is supported by near-bottom magnetic anomaly studies (e.g., Macdonald, 1977), in contrast to earlier continuously evolving models, which leaned heavily on magnetic anomaly patterns mapped at or above the sea surface. The small-scale topographic and magnetic structures occurring at oblique angles to regional spreading directions (e.g. Macdonald and Holcombe, 1978) are typical of the kind of complexity being recognized in these detailed studies. The study of small variations in basalt composition, particularly in trace-element distributions (White and Bryan, 1977), has led to the realization that mid-ocean ridge basalts cannot be interpreted as simple products of either fractional crystallization or partial melting, but that a complex interaction of these processes, complicated by episodic magma injection into evolving magma chambers (O'Hara, 1977), is more likely to have taken place.

The differences between fast-spreading ridges, (with no axial rift), and slow-spreading ridges (with rifts) have been illuminated by seismic refraction studies that show much stronger evidence of active magma chambers at fast-spreading ridges than at slowly spreading ridges (Orcutt, et al., 1975; Fowler, 1977). It has even been suggested that magma bodies may only be episodically present (Tapponnier and Francheteau, 1978) at slowly spreading ridges, but the geochemical data do not support this interpretation (Walker, et al., in press). Thermal calculations (Sleep, 1975)

suggesting episodic magma chambers along slowly spreading ridges are therefore unlikely to be correct. Clearly, there is some minimum spreading rate below which the cooling effect of circulating sea water will overwhelm the heat supplied by new magma, but this does not seem to have happened along most of the present ridge system.

The hot-spot (i.e. non-plate margin-type) volcanism that occurs in places like the Afar, Iceland and the Azores will be considered in a later section because we interpret it not so much as an anomalous type of divergent plate-boundary volcanism but as a normal part of the wide spectrum of hot-spot volcanism. Divergent plate boundaries have developed across the sites of these hot spots.

All but a tiny proportion of oceanic crust has been subducted--that is, permanently removed from the Earth's surface at least in any recognizably original form. Of the tiny sample that has escaped this fate, much is badly shredded and dismembered, particularly the material preserved in steeply-inclined zones and/or at deeper structural levels, even in Mesozoic and Tertiary orogenic belts. The general observation that recognizable ophiolite complexes become scarcer in older orogenic belts is explained by the greate uplift and erosion which have generally occurred there, compared with younger belts. Other reasons for their reported scarcity may include the past unfamiliarity of workers in older belts with the utility of the ophiolite concept, the possibility that major continental collision and suturing was generally more effective and intense further back in time, and the effect of the smaller length of older orogenic belts remaining exposed or preserved compared with younger ones. Well-described ophiolite complexes of Palaeozoic age are known from the Appalachians and Caledonides (Bird, et al., 197', 1978; Dewey and Bird, 1971; Church and Stevens, 1971; Sturt, et al., 1979), and less detailed descriptions of some of the

obviously widespread Palaeozoic complexes in the Urals and central Asian orogenic belts are available (for example, Abdulin, et al., 1974; Makarychev and Shtreys, 1973).

Definite examples of pre-Palaeozoic ophiolite complexes have been identified only in Pan-African orogenic belts in Morocco (LeBlanc, 1976) and Saudi Arabia (reported in Brown, 1978), and in the older Baikal orogenic belt in the U.S.S.R. (Klitin and Pavlova, 1974). Many other dismembered pieces surely remain to be identified in Proterozoic orogenic belts, since the well-studied belts, for example, the Labrador-Cape Smith-Nelson (Wilson, 1968b) and the Coronation orogenic belts (Hoffman, this volume), so clearly developed through rifting and later collision (Wilson cycles); dated dikes reveal that in both cases rifting began about 2.15 Ga ago. The suggestion (Burke, et al., 1976) that oceanic crustal samples (dismembered ophiolite complexes) should exist in Archean greenstone belts has not, to our knowledge, been confirmed. Because there is an extensive greenstone-belt terrane (the Birrimian of West Africa) of similar age to the Coronation orogen, it seems to us very likely that dismembered ophiolites will eventually be recognized in Archean greenstone belts. Since the Birrimian greenstones were formed at the same time that ocean opening and closing (which must have involved sea-floor spreading and subduction) is recorded in the Coronation and other orogens, samples of the oceanic crust of that time may be preserved in the Birrimian terrane. Since this terrane does not differ in any significant way from older Archean greenstone terranes, the inference that oceanic crust may be preserved in Archean terranes seems to us unexceptionable. Extensive areas of mafic volcanics of tholeiitic compositions appropriate to the ocean floor exist in the Archean greenstone belts, but equally widespread plutonic equivalents are generally lacking. Moores (1973) suggested that older Pre-Cambrian oceanic crust contained more anorthosite than younger material;

this has yet to be verified. It is possible that the extensive tholeiitic submarine lavas of the greenstone belts include, besides lava generated as ocean floor and island arcs, large thicknesses generated by intraplate flood-basalt type magmatism; this is discussed below in the section on intraplate volcanism.

VOLCANISM AT PLATE MARGINS: TRANSFORM MARGINS

Very much less volcanism occurs at transform boundaries than at convergent and divergent plate boundaries and all, or nearly all, of what does occur there is associated with local extensional areas, commonly termed "pull-aparts." "Leaky transforms" are most probably the same features, less well-defined because they are submarine. The young alkalic basalts dredged along the St. Paul and Romanche fracture zones are the only rocks presently known that are likely to have been erupted on oceanic transform/fracture zones away from the places where spreading ridge axes abut such zones. Even in these two cases, the tectonic details of their occurrence and the local abundance of the volcanics (or lack of it) are poorly known. It is also hard to prove that these volcanics are not related to areas of "hot-spot" activity, with relatively little associated volcanism, like the Jos Plateau and Air regions of the African plate. These areas are clearly identifiable since they are not submarine, lie within a plate and are obviously associated with a structurally and topographically defined uplift. The fact that there is an uplift at St. Paul's Rocks perhaps favors the idea that these volcanics are "hot-spot"-related rather than related to secondary extension across the transform/fracture zone.

Pull-apart basins along large strike-slip faults on land (e.g. the Salton Sea, the Dead Sea rift, near the western end of the Altyntagh Fault, and small basins along the North Anatolian fault) reveal that such pull-aparts are not common, that they are usually small relative to the extent of the

transform system along which they occur and that the volume of volcanics directly associated with the pull-apart is also rather minor. It may be that larger volumes of magma occur at depth and that the surface expression is not representative of the amount of magmatism in these pull-aparts; the geothermal activity in the Salton Sea area may be evidence of this.

It has been proposed that other volcanic rocks which occur in some places in the general region of large continental transform fault systems (e.g. the young Arabian basalts, the Hsing-An basalts of China, and some of the young volcanics in the vicinity of the San Andreas Fault) are in a less direct way related to the transform-zone tectonics. Since present evidence for a strong connection is unconvincing, we treat such volcanics under intraplate volcanism.

Perhaps because of the small volume of transform-related volcanics, and their susceptibility to later tectonic disruption, no well-documented examples are known from older, inactive transform zones. For example, well-studied pull-apart basins formed on an extensive, large-displacement Carboniferous strike-slip fault zone (probably an old transform system) through the Canadian Maritime Provinces (Belt, 1969) do not contain any known examples of contemporaneous volcanics. The exposed portion of the possibly correlative (Wilson, 1962) Great Glen Fault system does not show any obvious pull-apart structures. A much older example (around 1.8 Ga) of a preserved piece of large-displacement strike-slip fault, the McDonald Fault, also has no known volcanics associated with its movement, despite having very thick accumulations of coarse clastic sediments (P.F. Hoffman, pers. comm.), which probably were preserved by pull-apart tectonics.

VOLCANISM AT PLATE MARGINS: CONVERGENT MARGINS

In present-day arc systems andesitic volcanism dominates, although overall about 20% is basaltic (Ewart, 1976). The volcanic zone always lies above a subducting slab of oceanic lithosphere, generally where the latter is between 100 and 150 km deep. It is therefore inferred that the descending lithosphere is involved in production of arc magmas, although exactly how and where these magmas are produced is unresolved. While great variability exists in the detailed features of volcanic arcs, the most prominent contrast in volcanism is that silicic, large-volume ignimbrites with high Na and K content are concentrated in, if not confined to, areas underlain by continental-type lithosphere, even though, as in New Zealand and S. Alaska, this may consist of (geologically) recently accreted material. Some involvement of the lithosphere underlying these Andean-type volcanic arcs in the generation of their magmas seems likely.

Arcs built on oceanic crust tend to contain proportionately more mafic and less silicic volcanics, although precise estimates are difficult to make since most of these arcs are submerged, and preservation is biased against pyroclastic and clastic materials. Burke, et al. (1976) pointed out that the collision and accretion of this kind of arc, together with remnants of the intervening marginal basin floors and fill, will produce greenstone-granodiorite belt geology like that of the Archean Superior Province and similar terranes elsewhere. These include at least one that is of post-Archean age (the Birrimian of West Africa--1.8 Ga), and thus contemporary with orogenic belts clearly formed from the operation of the Wilson cycle. The present southwestern Pacific contains areas that we envisage as close analogs to those that formed the greenstone terranes. Andesites in present arcs do not have any significant geochemical differences from andesites in

Archean greenstone belts. As the overwhelming bulk of andesites is made at convergent margins today, possibly in response to the liberation of water from hydrated phases in the subducted slab at significant depths in the mantle, we see no good reason to suppose that they were made in any grossly different way in the past. Greenstone-type terranes are not unique to the Archean. We pointed out above (see also Burke and Dewey, 1972) the example of the Birrimian, and smaller analogs can be found in Palaeozoic and Mesozoic orogenic belts (Burke, et al., 1976). More extensive analogs probably exist, awaiting proper description, in the Palaeozoic orogenic area of Central Asia (Burke, et al., 1978a). The area of older orogens preserved intact from the tectonic effects of younger orogenies decreases with age. It is entirely possible that the sample we have of Archean orogenic terranes is not representative of their original proportions, due to the accidents of the siting of later rifts and, hence, collisions (Wilson cycles). Assuming, however, that the greater abundance of greenstone belts in the Archean reflects a real secular tectonic change, then it is reasonable to suppose that it is connected with the Earth's greater heat production in the past, which, as suggested above, resulted in faster plate motion and greater length of plate boundary. Such properties would most probably have led to more arc generation in a given time, thus accounting for the greater abundance of greenstone-type terranes in the past without excluding them from post-Archean terranes.

Plate convergence eventually leads to continental collisions. These, in recent examples (Fig. 3), give rise to volcanics very similar to those found in Andean arcs (Burke, et al., 1974; Kidd, 1975; Şengör and Kidd, 1979). Older examples of such calc-alkaline, K-rich volcanics and/or their subjacent post-kinematic granites are well known in collisional orogens up to about 2 Ga. old: e.g., the granites and other plutons of the Appalachian-Caledonian belt (Dewey and Kidd, 1974); the post-kinematic granites and

local rhyolites of the Pan African; the rhyolites of the St. Francois Mountains (Bickford and Mose, 1975) related to the Elsonian orogeny which we suggest is collision-induced; and the volcanics of the Bear Province (Hoffman, McGlynn and others, in prep.), related to collision about 1.8 Ga ago at the end of the Wilson cycle on the site of the Coronation orogen. Pieces of continental crust older than 1.8 Ga are either too small or too deeply uplifted and eroded to distinguish collisional from arc-related magmatic products with any confidence, but we see no reason to suppose their absence.

INTRAPLATE TYPE VOLCANISM: HOT SPOTS AND FLOOD BASALTS

Young volcanism not related to plate boundaries is widespread (Burke and Kidd, 1975). Despite the great variety in its expression, we feel that there are sufficient common elements among different instances of non-plate margin volcanism that it is inappropriate not to treat them in a single category. Many areas of active intraplate volcanism are well-exposed in Africa and have been well studied; the association in them of volcanics and a structurally and topographically high area is well-established (Burke and Whiteman, 1973; Thiessen et al., 1979). The volcanics of Dakar, near sea level, are not an exception, since subsurface data (Spengler and Deteil, 1966) reveal an underlying youthful structural elevation. There is a complete spectrum of intraplate volcanic areas in Africa, from uplifts without any volcanism (e.g. Fouta Djallon) through minor volcanism of alkaline type (e.g. Air) to more abundant alkaline volcanism (e.g. Tibesti), to areas of voluminous tholeiitic flood basalts together with comparatively minor alkaline volcanics (e.g. Ethiopia). In all cases a structural uplift is associated with the volcanic area; the size of this uplift varies, although most tend to be 100 to 200 km across (Burke and Whiteman, 1973). There

seems to be a tendency for the areas with larger volumes of volcanics to have larger diameters of uplifts, although it becomes possible to resolve subsidiary uplifts within the larger ones (e.g. Ahaggar; Black and Girod, 1970). Thus it is not wholly clear whether the larger uplift structures are merely groups of smaller ones, or whether the smaller diameter uplifts in them are secondary to the larger, but generally lower amplitude uplifts.

This spectrum of intraplate volcanism can be seen, less well displayed, on other continents, but it is more significant that it is also developed in the oceans, and that the alkaline and tholeiitic volcanics in oceanic hot spots are essentially indistinguishable from those in continental hot spots. Again, the African plate provides many of the best examples. Oceanic intraplate volcanic areas also show a variation in volume from the largest, like Hawaii, with abundant tholeiite and minor alkaline volcanics, to relatively small edifices, like Ascension, that are mainly alkaline volcanics. Intraplate volcanic areas of very small volumes, like some of those on the African continent, are harder to detect in the oceans, and uplifts without accompanying volcanism harder still. Nevertheless, careful study has detected them in one area (Menard, 1973).

Because alkaline magmas are erupted in places within present island arcs and collisional orogens, it is not possible to detect "intraplate type" or "hot-spot" type activity unambiguously within such zones of convergence even though it may well occur. However, it is possible to detect such volcanism along divergent plate boundaries because of the alkaline character of some of the magmas, the excess volume of magma and the associated structural uplift. The occurrence of "hot-spot" type volcanism at discrete and long-lasting sites along divergent plate boundaries is, we suggest, one of the most significant properties of their distribution, and helps considerably in

winnowing the many hypotheses put forward for their origin. In particular, since the lithosphere is thin to virtually non-existent at spreading ridge crests and is created progressively away from them, it is unlikely that either the crack propagation hypotheses of Turcotte and Oxburgh (1973) and Oxburgh and Turcotte (1974), or the dense anchor-asperity hypothesis of Shaw and Jackson (1973) are correct. An "active mantle" hypothesis for hot-spot origin, such as proposed by Wilson (1963), is more satisfactory.

A similar spectrum of size and volume of magmas can be seen in those hot spots located on divergent plate boundaries. The central and north Atlantic contains the best examples. These range from the anomalously elevated area at 45°N on the ridge, where there seems to be relatively little excess volcanism, through the Azores, with mostly alkaline volcanism constructing small islands above sea level, to the extreme of Iceland, with its voluminous excess tholeiitic volcanism and, in relative terms, minor alkaline magmatism. Anomalous (with respect to spreading-ridge basalt) geochemical signatures, particularly $^{87}Sr/^{86}Sr$ (White, et al., 1976) correlate exactly with discrete anomalously elevated ridge crest areas, in Iceland, 45°N, the Azores, and near 34°N (Colorado Seamount). The distinctive geochemistry of the "excess" magmas argues strongly for a separate (deeper) source for them (Schilling, 1973) compared with normal spreading-ridge basalts. It is because the latter are usually produced in a fairly constant amount through a wide range of spreading rates that the "excess" character of the hot-spot magmatism along ridge axes can be easily detected.

The connection between flood basalts and the largest hot spots is clearly shown by Iceland. The island itself consists of flood basalts, and the hot-spot tracks that lead away from Iceland go to large areas of flood basalts in East Greenland and the northern British Isles (Fig. 4). These basalts were erupted during the initial rifting that lead to successful

sea-floor spreading in the northern Atlantic beginning about 60 Ma ago. The relics of associated central alkaline volcanoes of that age are well known from Scotland. A younger but similar situation is seen in the Afar (Fig. 3), where the Ethiopian Traps erupted before and up to the opening of the Red Sea and Gulf of Aden. If the Pacific plate were not moving so rapidly with respect to the source of the Hawaiian magmas, the accumulation of igneous material at present rates of production would clearly rival Iceland, and although the Hawaiian Emperor hot-spot trace (Fig. 4) does not lead back to a site of rifting and a large flood basalt pile, it is appropriately grouped with other flood basalt-producing objects. Several other large hot spots do have tracks leading back to rifting sites and accumulations of flood basalts (Fig. 4), in particular Tristan/Gough to the Kaokoveld and Parana basalts which were erupted just before and up to the opening of the South Atlantic 120 Ma ago; and Reunion to the Deccan Traps, which were erupted about 65 m.y. ago just before rifting in the Gulf of Khambat and removal of the Seychelles from India (McKenzie and Sclater, 1971). The Galapagos hot spot, although its tracks do not lead back to preserved flood basalts, is sited on a spreading ridge that started about 25 Ga ago by rifting across older oceanic crust generated at the East Pacific Rise (Hey, 1977). It may not be a coincidence that the Galapagos hot spot, as shown by its area and the volume of magmatic products in its tracks, is a relatively large one, like others that have been listed as associated with flood basalt production and initial ocean opening. Ocean opening in the Labrador Sea-Baffin Bay was accompanied by the flood basalt volcanism (about 60 Ma) which is now preserved on Disko Island, and Cape Dyer of Baffin Island. The hot spot that produced this volcanism made the shallow sill to Davis Strait as a track, but is now obviously extinct.

Perhaps it is also not a coincidence that spreading ceased in the Labrador Sea-Baffin Bay at the same time or shortly after the hot spot died. This case illustrates the point emphasized by Vogt (1972), that the volumes of magma generated in hot-spot sites vary with time, although, unlike him, we do not think there is strong evidence for synchroneity in this variation among many hot spots. Thus, the amount of magma produced by the Hawaiian hot spot, judged by the volume of the track, was very small at the time of and shortly after the bend formed in the Hawaiian-Emperor chain. Its volume may have been no more impressive at that time than one of the smaller present ocean island hot spots, like St. Helena. The point we wish to make here is that voluminous flood basalt-producing hot-spots are the same kind of object as the smaller ones, and that one can change into the other, and back again, with time. They may also die out, in terms of their volcanic expression, temporarily or permanently. They vary a great deal in the length of time during which large quantities of tholeiitic magma are produced, from the short burst of the Columbia River basalts (Baksi and Watkins, 1973) and Deccan Traps (Wellman and McElhinny, 1970) to the more extended histories of Iceland and Hawaii.

Large flood tholeiite events in the oceans seem, in some instances, to produce huge sill complexes, probably because of the limited abilities of even mafic lava to travel underwater before chilling. Such sill complexes (Fig. 4) have been identified underlying large portions of the Caribbean (Burke, et al., 1978b) and the submarine plateaus of the western Pacific (Winterer, 1976). The Mid-Pacific Mountains, one of the latter, may perhaps be traced to the hot spots of either Easter Island and/or Pitcairn Island through the Tuamotu-Line Islands track. Sill complexes are, of course, also important components of many continental flood basalt

events, particularly the early Jurassic Karoo and Ferrar dolerites of South Africa and the Antarctic, and the mid-Cretaceous Isachsen diabase of the Canadian Arctic (Fig. 4).

It is of note that all major flood basalt events (where at least some of the extensive lavas and/or sills are still preserved) were connected with extensive rifting and, in most cases, with successful opening of an ocean. Two large and well known ones not associated with successful ocean opening are the Columbia River basalts and the Siberian Traps.

Well-preserved remnants of older flood basalt and/or sill complexes are seen back to 2.15 Ga. Most remnants are found within old rifts and aulacogens because the accidents of erosion through geological time have claimed any more extensive basalts lying relatively higher outside them. The most prominent flood basalt remnants of Palaeozoic or older age lying outside old rifts and of any significant extent are the 600 Ma-old Antrim basalts of N.W. Australia, the 1100 Ma-old Keweenawan lavas and sills north of Lake Superior, and the 1900 Ma- old Kimberly Plateau basalts and sills, also in N.W. Australia. Within old rifts there are many more examples; prominent ones in North America are the rifts of about 1100 Ma, which include the Coppermine River (N.W. Territories), part of which may be outside the rift defined by Burke and Dewey (1973), Seal Lake (Labrador), Gardar (S. Greenland), Keweenawan and its extension in the mid-continent rift and gravity high, and various aulacogens along the Cordilleran margin of the North American craton, including one exposed in part in the Grand Canyon. Another prominent rifting episode is that shown by the extensive tholeiitic sill complex (about 2.15 Ga old) that invades what we interpret as the initial rifting facies clastics and volcanics (arkoses, sandstones and basalts, etc. of the Seward and part of Attikamagen Formations) which are preserved in the fold and thrust belt of the Labrador Trough (Baragar, 1967).

Associated with this 2.15 Ga rifting and ocean opening event is an extensive dike swarm (Fahrig and Wanless, 1963; Stevenson, 1968) that runs (in Archaen basement) sub parallel with the northern end of the Labrador Trough and then turns to run obliquely along the Cape Smith belt, the continuation of the Labrador Trough in northern Ungava. This dike swarm is the last relic of a flood basalt event; a very similar although smaller example is seen close to and subparallel with the western margin of the Appalachians in northern Newfoundland (Williams, 1967), where it (and nearby small relics of flood basalts) was associated with the early Cambrian or slightly older opening of the ocean Tuzo Wilson called the Proto-Atlantic (1966), now termed the Appalachian Ocean or Iapetus. Dike swarms as relics of flood basalt and rifting events are common. The MacKenzie dikes, which extend a great distance NNW across the western part of the Canadian shield are of the same age as the Keweenawan rifting and flood basalt event. There are no well-preserved segments of continental rifted margins older than 2.15 Ga, the opening age of the Labrador Trough and Coronation oceans. However, there are examples of extensive tholeiitic dike swarms, from which we infer that rifting events like those recorded in younger rocks took place. The Ameralik dikes in West Greenland (McGregor, 1973), which are at least 3.0 Ga and perhaps as much as 3.6 Ga old (Pankhurst, et al., 1973) are evidence of the oldest rifting event directly recorded. The nature of the older rocks they cut, containing calc-alkaline volcanics, is to us indirect evidence of subduction, sea-floor spreading and yet older rifting.

One feature of low-metamorphic grade Archean volcanic terranes is that they do not--with very minor exceptions which could be younger than Archean (Cooke and Moorhouse, 1969)--contain alkaline volcanic and plutonic rocks like those which have been emplaced at present hot spot sites and are preserved sparingly in rocks up to about 2.5 Ga old. Judging by the largest

present hot spots, perhaps most of them were large in the Archean and erupted huge quantities of flood basalt-type tholeiite. Lavas and sills generated in this way, if preserved from subduction, might be represented within the extensive basalts in greenstone belts. Alternatively, efficient subduction of Archean oceanic lithosphere may have removed essentially all the hot spot material and may have preferentially preserved island arc edifices, which even today contain only rare occurrences of alkalic volcanics. Added to the small area of preserved Archean rocks, this small chance of preservation of alkalic volcanics may explain why they are so rare in such terranes.

FLUCTUATION IN VOLCANIC ACTIVITY?

An important question is whether, and if so by how much, the amount of volcanism on Earth has fluctuated with time. This question is fraught with difficulties. Older rocks are generally preserved in smaller proportions relative to younger ones, so that the occurrence of volcanic rocks per unit time could be expected to be less. Long ago, however, since heat generation was greater, volcanism may also have been greater, and these effects may partly cancel. Because oceanic rocks are destroyed by subduction and obduction, the record is necessarily incomplete; generally only arc and continental rocks are preserved. In spite of these difficulties, some workers have discerned episodicity in igneous activity, especially in the Precambrian. The crude technique, popular twenty years ago, of plotting histograms of isotopic ages has fallen into disuse with the recognition of the complexity of the record. A particularly significant observation is that the proportions of continental area representative of a particular time interval and available for study are so small. For example, the Superior Province, which is the largest Archean area on Earth, representing about half the

total area of Archean rocks not obscured by later events, amounts to only 1% of continental area. A more sophisticated approach considering the isotopic compositions of Nd, Sr and Rb has been used by several authors (e.g. McCulloch and Wasserberg, 1978) to show that many Precambrian continental rocks became isolated from the main mantle resevoir during an interval of about 200 Ma roughly 2.7 to 2.5 Ga ago. This is a most interesting and unexpected result.

Turcotte and Burke (1978) used an indirect method to estimate volcanic fluctuation with time during the Phanerozoic. Realizing that sea level responds to the volume of the mid-ocean ridges, they inferred that the times when the continents were most flooded were times when the ridges were most active. By crudely calculating the proportions of heat escaping through conduction and ocean-floor aging, they were able to estimate that nearly twice as much heat was escaping the Earth through the cooling of aging ocean floor during the late Cretaceous episode of continental flooding as is escaping in this way now. In order to keep the Earth at roughly the same volume, they inferred that plate consumption also peaked at this time, which is consistent with the familiar high concentration of circum-Pacific batholithic emplacement during the Late Cretaceous and the Late Cretaceous concentration of emplacements of Tethyan ophiolites.

CONCLUSION

It has been impossible to cover all aspects of volcanism through time in such a short review, but it seems appropriate to point out that the history of volcanic activity on Earth as summarized here is best viewed as Tuzo Wilson (1968a) first proposed, in the context of cycles of ocean opening and closing. Not only can rift, plate-margin and collisional

volcanism be well accommodated within the framework of "Wilson Cycles" (Dewey and Burke, 1974), but so can the role of hot spot (intraplate type) volcanism, especially of the Hawaiian kind (Wilson, 1963).

REFERENCES

Abdulin, A.A., Bespalov, V.F. and Shelepova, T.N., 1974, Ophiolite belts of Kazakhstan, Doklady Akad. Nauk. SSSR. 218, 159-162 (20-22 in translation)

Baksi, A.K. and Watkins, N.D., 1973, Volcanic production rates: comparison of oceanic ridges, islands and the Columbia River basalts, Science, 180, 493-496.

Ballard, R.D. and van Andel, T.H., 1977, Morphology and tectonics of the inner rift valley at 36°50'N on the Mid-Atlantic Ridge, Geol. Soc. Amer. Bull., 88, 507-530.

Baragar, W.R.A., 1967, Wakuach Lake Map area, Quebec-Labrador (230), Geol. Surv. Canada, Mem. 344.

Belt, E.S., 1969, Newfoundland Carboniferous stratigraphy and its relation to the Maritimes and Ireland. pp. 734-753, in M. Kay (ed.) North Atlantic-Geology and Continental Drift, Amer. Assoc. Petrol. Geologists, Mem. 12.

Bickford, M.E. and Mose, D.G., 1975, Geochronology of Precambrian rocks in the St. Francois Mountains, southeastern Missouri, Geol. Soc. Amer. Spec. Pap. 165; 48 pp.

Bird, J.M., Dewey, J.F. and Kidd, W.S.F., 1971. Proto-Atlantic ocean crust and mantle: Appalachian/Caledonian ophiolites, Nature, Phys. Sci., 231, 28-31.

Bird, J.M., Dewey, J.F. and Kidd, W.S.F., 1978, The Mings Bight ophiolite complex, Newfoundland: Appalachian oceanic crust and mantle, Can. J. Earth Sci., 15, 781-804.

Black, R. and Girod, M., 1970. Late Palaeozoic to recent igneous activity in West Africa and its relationship to basement structure, pp. 185-210 in African Magmatism and Tectonics, T.N. Clifford and I.G. Gass (eds.) Oliver and Boyd, Edinburgh.

Brown, G.F., 1978. Arabian-Nubian shield discussed on the spot. Geotimes, 23 (7) p. 24.

Bryan, W.B. and Moore, J.G., 1977. Compositional variations of young basalts in the Mid-Atlantic Ridge rift valley near 36°49'N, Geol. Soc. Amer. Bull., 88, 556-570.

Burke, K.C.A. and Dewey, J.F., 1972. Orogeny in Africa, pp. 583-608 in Dessauvagie T.F. and Whiteman, A.J.,(eds). African Geology Univ. Ibadan Press.

Burke, K. and Dewey, J.F., 1973. Plume-generated triple junctions: key indicators in applying plate-tectonics to old rocks. Jour. Geol. 81, 406-433.

Burke, K.C. and Kidd, W.S.F., 1975. Earth, heat flow in pp. 165-169 in McGraw-Hill Yearbook of Science and Technology, McGraw-Hill, New York.

Burke, K. and Kidd, W.S.F., 1978. Were Archean continental geothermal gradients much steeper than those of today? Nature, 272, 240-241.

Burke, K. and Whiteman, A.J., 1973. Uplift, rifting and the breakup of Africa pp. 735-755 in, Tarling, D.H. and Runcorn, S.K. (eds.) Implications of Continental Drift to the Earth Sciences, vol. 2, Academic Press, London.

Burke, K., Dewey, J.F. and Kidd, W.S.F., 1974. The Tibetan Plateau: its significance for tectonics and petrology, Geol. Soc. Amer. Abstr. with Progr. 6, 1027-1028.

Burke, K., Dewey, J.F. and Kidd, W.S.F., 1976. Dominance of horizontal movements, arc and microcontinental collisions during the later permobile regime, pp. 113-129, in B.F. Windley (ed.) The Early History of the Earth, Wiley, London.

Burke, K.C.A., Dewey, J.F. and Kidd, W.S.F., 1977. World distribution of sutures - the sites of former oceans, Tectonophysics, 40, 69-99.

Burke, K.C.A., Delano, L., Dewey, J.F., Edelstein, A., Kidd, W.S.F., Nelson, K.D., Şengör, A.M.C. and Stroup, J., 1978a. Rifts and Sutures of the World, Contract Rept. NAS5-24094. Geophysics Branch ESA Division Goddard Space Flight Center, Greenbelt, MD 238 pp.

Burke, K., Fox, P.J. and Sengor, A.M.C., 1978b. Buoyant ocean floor and the evolution of the Caribbean, J. Geophys. Res., 83, 3949-3954.

Church, W.R. and Stevens, R.K., 1971. Early Palaeozoic ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences, J. Geophys. Res., 76, 1460-1466.

Cooke, D.L. and Moorhouse, W.W., 1969. Timiskaming volcanism in the Kirkland Lake area, Ontario, Canada, Can. J. Earth Sci., 6, 117-132.

Dewey, J.F. and Bird, J.M., 1971. Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland. J. Geophys. Res. 76, 3179-3206.

Dewey, J.F. and Burke, K.C.A., 1973. Tibetan, Variscan and Precambrian basement reactivation: products of continental collision, J. Geol. 81, 683-692.

Dewey, J.F. and Burke, K., 1974. Hot spots and continental break-up: implications for collisional orogeny, Geology, 2, 57-60.

Dewey, J.F. and Kidd, W.S.F., 1974. Continental collisions in the Appalachian-Caledonian orogenic belt: variations related to complete and incomplete suturing, Geology, 2, 543-546.

Dewey, J.F. and Kidd, W.S.F., 1977. Geometry of plate accretion, Geol. Soc. Amer. Bull. 88, 960-968.

Ewart, A., 1976. Mineralogy and chemistry of modern orogenic lavas - some statistics and implications, Earth Planet. Sci. Lett., 31, 417-432.

Fahrig, W.F. and Wanless, R.K., 1963. Age and significance of diabase dyke-swarms of the Canadian shield, Nature, 200, pp. 934-937.

Fowler, C.M.R., 1977. Crustal structure of the mid-Atlantic ridge crest at 37°N, Geophys. J. Roy Astron. Soc., 47, 459-491.

Gansser, A., 1964. The Geology of the Himalayas, Wiley-Interscience, New York, 289 pp.

Hey, R., 1977. Tectonic evolution of the Cocos-Nazca spreading center, Geol. Soc. Amer. Bull., 88, 1404-1420.

Hoffman, P., this volume, Precambrian Wilson Cycle.

Kidd, W.S.F., 1975. Widespread late Neogene and Quaternary calc-alkaline volcanism on the Tibetan Plateau, EOS, Trans. Amer. Geophys. Union, 56, 453.

Klitin, K.A. and Pavlova, T.G., 1974. Ophiolite complex of the Baikal fold zone, Doklady Akad. Nauk SSR, 215, 413-416, (33-36 in translation).

LeBlanc, M., 1976. Oceanic crust at Bou Azzer, Nature,261, 34-35.

Lee, W.H.K., 1967. Thermal history of the Earth, Unpub. Ph.D. thesis, University of California, Los Angeles.

MacDonald, K.C., 1977. Near-bottom magnetic anomalies, asymmetric spreading, oblique spreading, and tectonics of the Mid-Atlantic Ridge near 37°N, Geol. Soc. Amer. Bull., 88, 541-555.

MacDonald, K.C. and Holcombe, T.L., 1978. Inversion of magnetic anomalies and sea-floor spreading in the Cayman Trough, Earth Planet. Sci. Lett., 40, 407-414.

McCulloch, M.T. and Wasserburg, G.T., 1978. Sm-Nd and Rb-Sr chronology of continental crust formation, Science, 200, 1003-1011.

McGregor, V.R., 1973. The early pre-Cambrian gneisses of the Godthaab district, West Greenland, Phil. Trans. Roy. Soc. London, A 273, 343-358.

McKenzie, D. and Sclater, T.G., 1971. The evolution of the Indian Ocean since the late Cretaceous, Geophys. J. Roy. Astr. Soc. 24, 437-528.

Makarychev, G.I. and Shtreys, N.A., 1973. Tectonic position of ophiolites of the southern Tien Shan., Doklady Akad. Nauk. SSSR, 210, 1164-1166 (92-93 in translation).

Mauk, F.T. and Johnston, M.J.S., 1973. On the triggering of volcanic eruptions by Earth tides, J. Geophys. Res. 78, 3356-3362.

Menard, H.W., 1973. Epeirogeny and plate tectonics, EOS, Trans. Amer. Geophys. Union, 54, 1244-1255.

Moores, E.M., 1973. Plate tectonic significance of Alpine peridotite types, pp. 963-975 in Tarling, D.H. and Runcorn, S.K., Implications of Continental Drift to the Earth Sciences, Academic Press, London, Vol. 2.

Morabito, L.A., Synnot, S.P., Kupferman, P.N. and Collins, S.A., 1979, Discovery of currently active extraterrestrial volcanism, Science, 204, 972.

O'Hara, M.J., 1977, Geochemical evolution during fractional crystallization of a periodically refilled magma chamber, Nature, 266, 503-507.

Orcutt, J., Kennett, B., Dorman, L. and Prothero, W., 1975. A low-velocity zone underlying a fast-spreading rise crest, Nature, 256, 475-476.

Oxburgh, E.R. and Turcotte, D.L., 1974. Membrane tectonics and the east African rift, Earth. Planet. Sci. Lett., 22, 133-140.

Pankhurst, R.J., Moorbath, S. and McGregor, V.R., 1973. Late event in the geological evolution of the Godthaab district, West Greenland, Nature Phys. Sci., 243, 24-26.

Peale, S.J., Cassen, P. and Reynolds, R.T., 1979. Melting of Io by tidal dissipation, Science, 203, 892-894.

Ramberg, I.B. and van Andel, T.H., 1977. Morphology and tectonic evolution of the rift valley at 36°30'N, Mid-Atlantic Ridge, Geol. Soc. Amer. Bull., 88, 577-586.

Schilling, J.G., 1973. Iceland mantle plume, Nature, 246, 141-143.

Şengor, A.M.C. and Kidd, W.S.F., 1979. Post-collisional tectonics of the Turkisn-Iranian plateau and a comparison with Tibet, Tectonophysics, 55, 361-376.

Shaw, H.R. and Jackson, E.D., 1973. Linear island chains in the Pacific: result of thermal plumes or gravitational anchors? J. Geophys. Res. 78, 8634-8652.

Sleep, N.H., 1975. Formation of oceanic crust: some thermal constraints. J. Geophys. Res., 80, 4037-4042.

Spengler, A. de E. and Delteil, J.R., 1966. Le bassin secondaire-tertiare de Cote d'Ivoire, pp. 99-113 in Du Reyre (ed.) Bassins sedimentaires du littoral africain 1e partie. Assoc. Serv. Geol. Afr. Paris.

Stevenson, I.M., 1968. A geological reconnaissance of Leaf River map area, New Quebec and Northwest Territories, Geol. Surv. Canada Mem. 356.

Sturt, B.A., Thon, A. and Furnes, H., 1979. The Karmoy ophiolite, southwest Norway, Geology, 7, 316-320.

Tapponnier, P. and Francheteau, J., 1978. Necking of the lithosphere and the mechanics of slowly accreting plate boundaries, J. Geophys. Res., 83, 3955-3970.

Thiessen, R., Burke, K. and Kidd, W., 1979. African hotspots and their relation to the underlying mantle, Geology, 7, 263-266.

Turcotte, D.L. and Burke, K.C.A., 1978. Global sea-level changes and the thermal structure of the Earth. Earth Planet. Sci. Lett. 41, 341-346.

Turcotte, D.L. and Oxburgh, E.R., 1973. Mid-plate tectonics, Nature 244, 337-339.

Vogt, P.R., 1972. Evidence for global synchronism in mantle plume convection and possible significance for geology, Nature, 240, 338-342.

Walker, D., Shibata, T. and DeLong, S.E. (in press), Abyssal tholeiites from the Oceanographer Fracture Zone, Contrib. Mineral. Petrol.

Wellman, P. and McElhinny, M.W., 1970. K-Ar age of the Deccan Traps, India, Nature, 227, 595-596.

White, W.M. and Bryan, W.B., 1977. Sr-isotope, K, Rb, Cs, Sr, Ba and rare-earth geochemistry of basalts from the FAMOUS area, Geol. Soc. Amer. Bull. 88, 571-576.

White, W.M., Schilling J-G, and Hart, S.R., 1976. Evidence for the Azores mantle plume from strontium isotope geochemistry of the central North Atlantic, Nature 263, 659-663.

Williams, H., 1967. Geological Map of Newfoundland 1:1m, Geol. Surv. Canada Map 1231A.

Wilson, J.T., 1962. Cabot Fault, an Appalachian equivalent of the San Andreas and Great Glen Faults and some implications for continental displacement. Nature, 195, 135-138.

Wilson, J.T., 1963. A possible origin of the Hawaiian Islands, Can. J. Phys., 41, 863-870.

Wilson, J.T., 1965. A new class of faults and their bearing on continental drift, Nature, 207, 343-348.

Wilson, J.T., 1966. Did the Atlantic close and then reopen? Nature, 211, 676-681.

Wilson, J.T., 1968a. Static or mobile Earth: the current scientific revolution, Proc. Amer. Phil. Soc., 112, 309-320.

Wilson, J.T., 1968b. Comparison of the Hudson Bay arc with some other features, pp. 1015-1033 in C.S. Beals and D.A. Shemstone (eds.) Science, History and Hudson Bay. Dept. Energy, Mines and Resources, Ottawa.

Winterer, E.L., 1976. Anomalies in the tectonic evolution of the Pacific pp. 269-278 in The geophysics of the Pacific Ocean Basin and its margin, Amer. Geophys. Union Monograph 19, Washington.

FIGURE CAPTIONS

Figure 1. Heat production in the Earth from decay of radioactive isotopes of U, Th and K through geological time (after Lee, 1967)

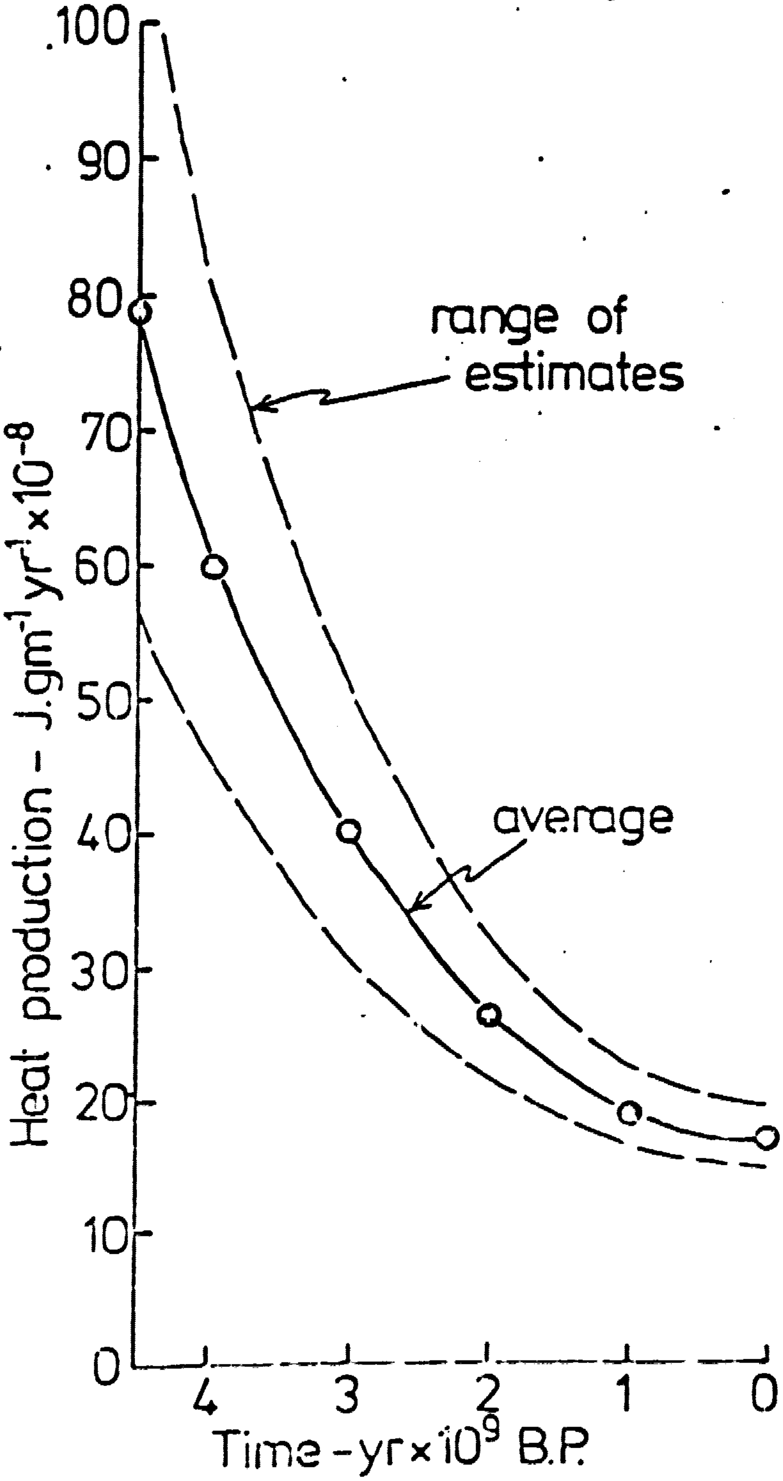
Figure 2. Cross-sections of oceanic spreading ridges (after Dewey and Kidd, 1977)

(a) Rifted, slow-spreading ridge.

(b) Non-rifted, fast-spreading rise crest.

Figure 3. Tectonic sketch map of the Tibetan Plateau and surrounding regions, emphasizing the Neogene-Recent volcanic rocks (stippled). Lines with black triangles -- active thrust boundaries; line with open triangles -- inactive thrust boundary; dot-dash lines -- active transcurrent faults.

Figure 4. World map with plate boundaries showing larger active hot spots (black circles); their tracks, i.e. volcanic accumulations left behind hot spots on moving plates (stipple); large areas of flood basalts and associated sills (horizontal ruling, dashed where basalts mostly subsurface); submarine equivalent of flood basalts, mostly sill complexes (vertical ruling - dashed where tentative). Approximate ages of flood basalts and sills given in millions of years, as are the ages of the oldest portions of some hot-spot tracks. Number underlined is age of youngest part of Davis Strait hot-spot track.



Heat production - J.gm^{-1}yr^{-1}yr^{-1} × 10^{-8}
100
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range of estimates
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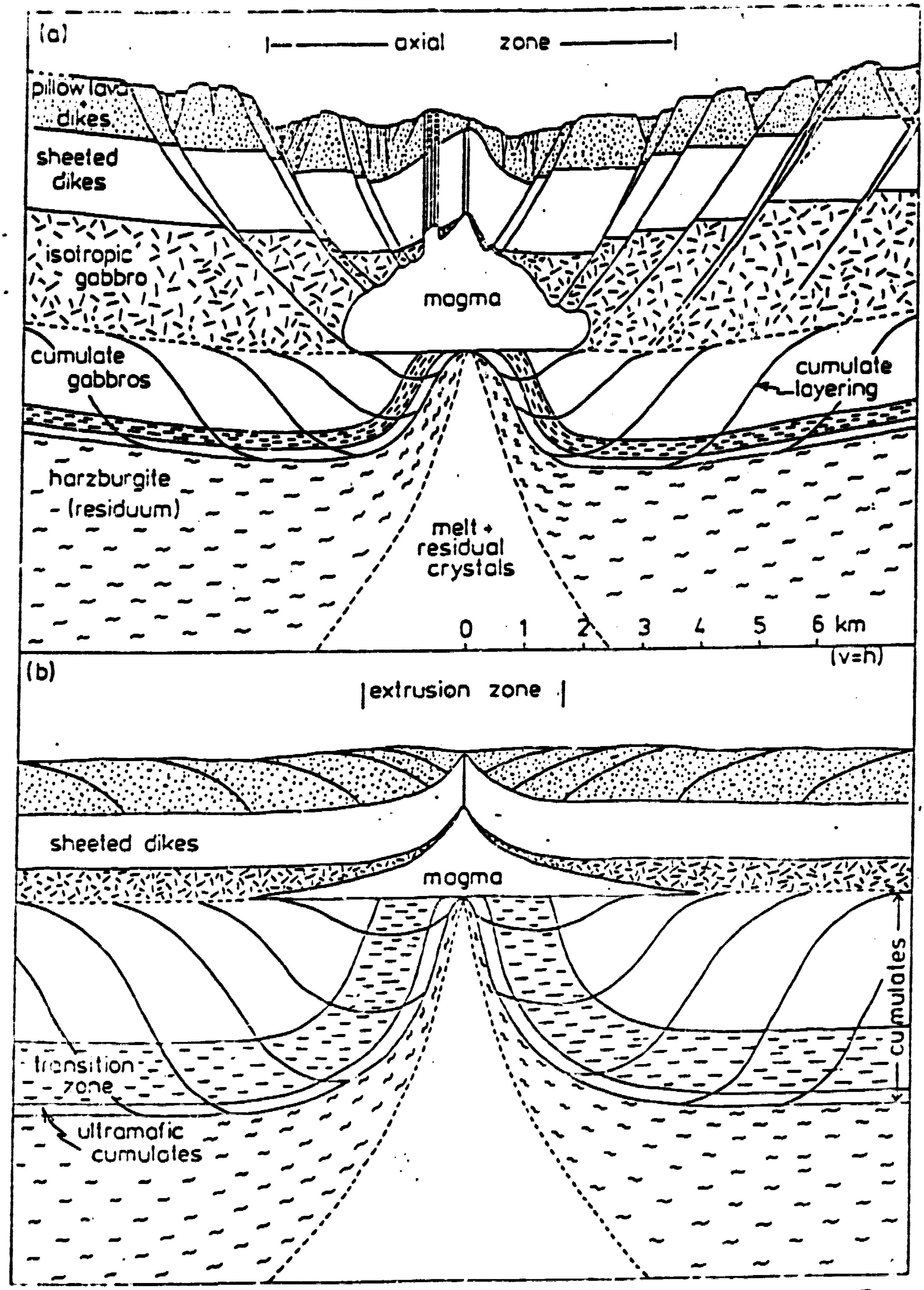
Time - yr × 10^{9} B.P.

Fig 1



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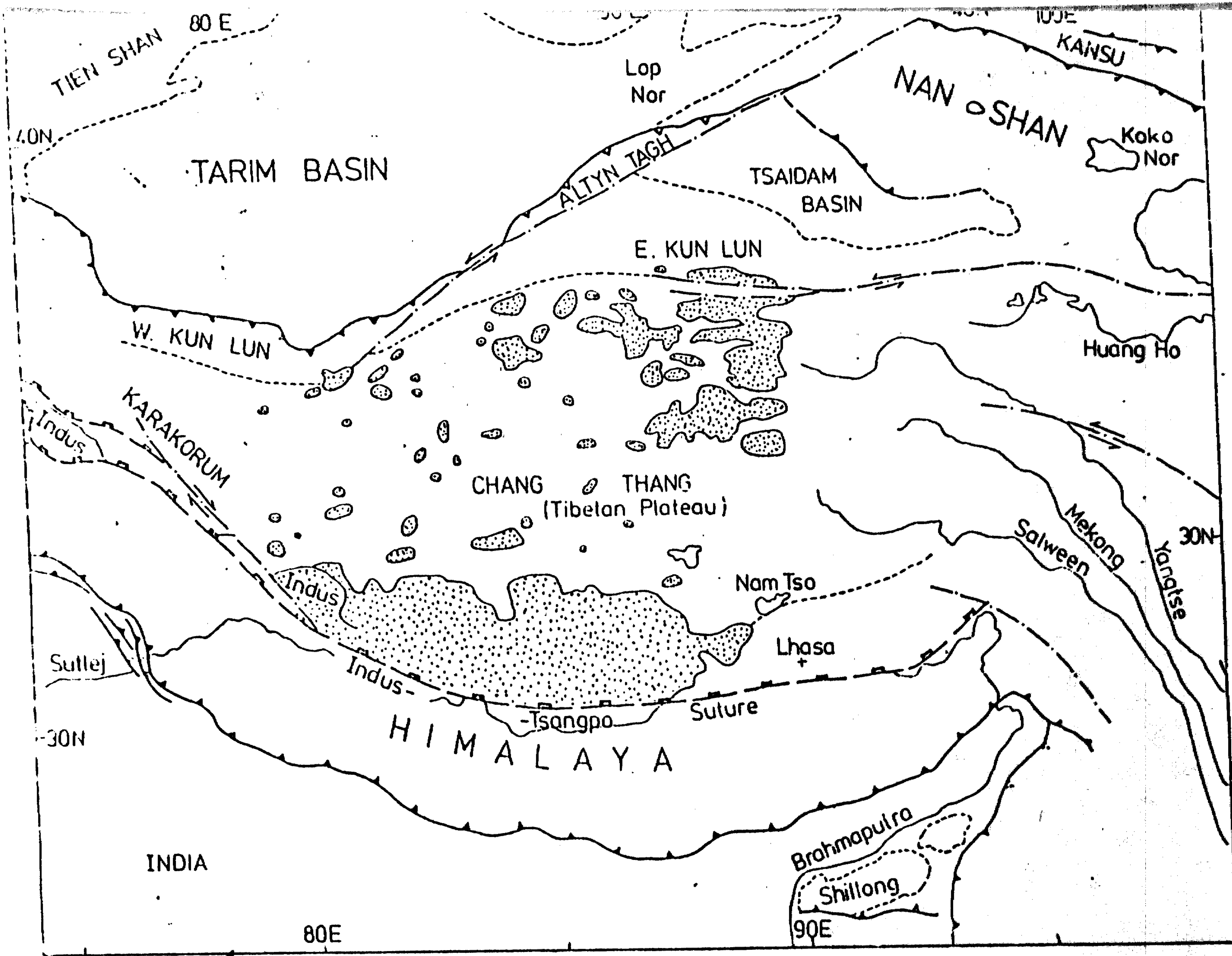
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NAN SHAN
KANSU
Koko
Nor
TSAIDAM
BASIN
E. KUN LUN
W. KUN LUN
Huang Ho
KARAKORUM
Indus
CHANG THANG
(Tibetan Plateau)
Mekong
Salween
Yangtse
30N
Nam Tso
Lhasa
Sutlej
Indus-
-Tsangpo
Suture
30N
HIMALAYA
INDIA
Brahmaputra
Shillong
80E
90E

Fig 3

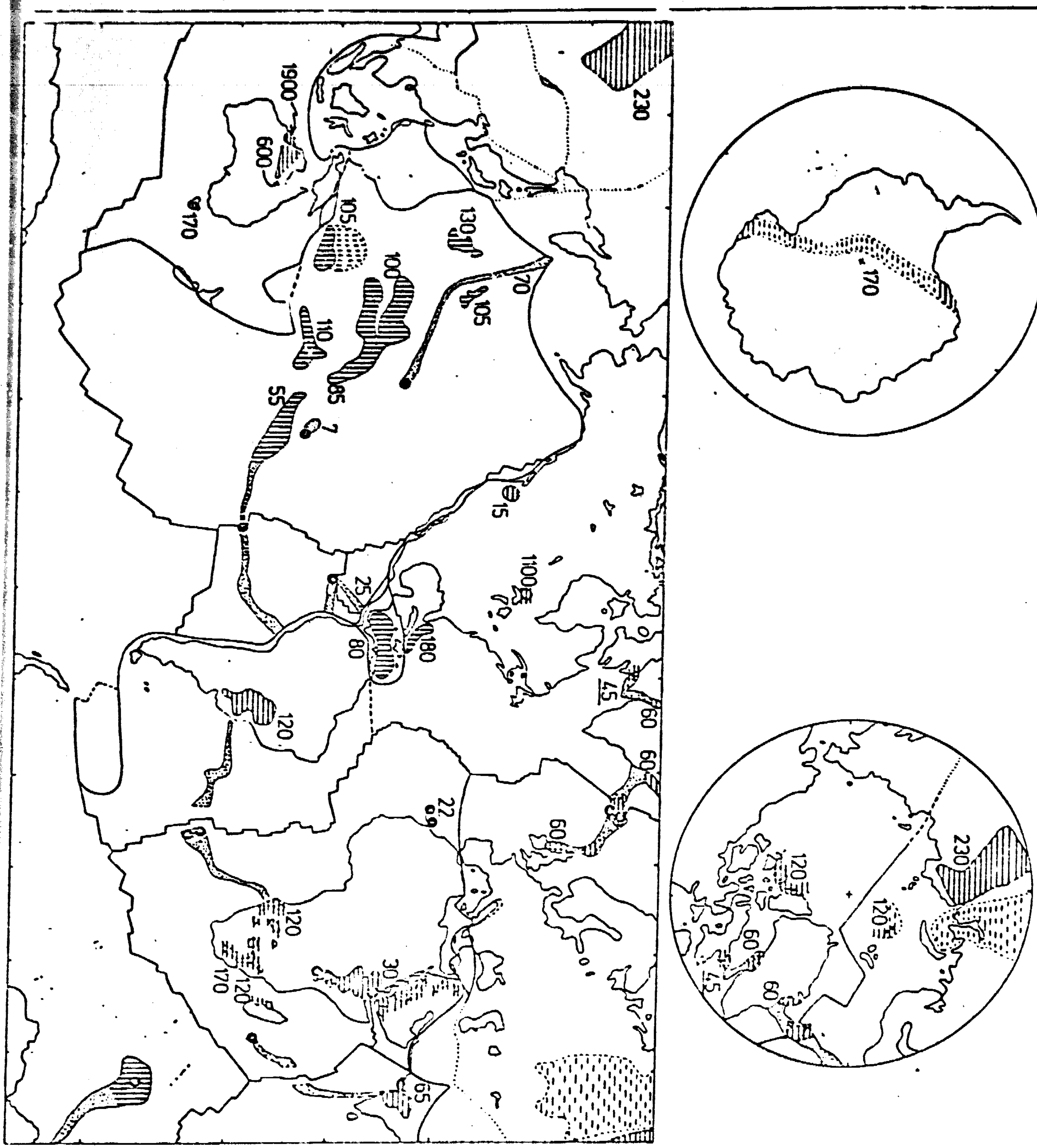
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Fig. 4.

Enclosure 3 43

ALBANY GLOBAL TECTONIC GROUP

Studies in Anomalous Continental Topography

#1 South America

by

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Albany, NY

1979

Introduction

Anomalous topography is relief which is not associated with either plate boundaries or normal non-plate margin edges of continents. (Burke, et.al., 1977) Erosional processes tend to reduce relief on land to near sea level within a few million years. Thus "normal" non-plate boundary areas lie within a few hundred meters of sea level. Anomalous areas considered here are those areas above 500 meters.

The two areas which fall into the anomalous category in South America are the Guyana Highlands and the Brazilian Highlands. The Guyana Highlands, in southern Venezuela and western British Guiana, constitutes the smaller, more equidimensional uplift. The Brazilian Highlands occupy a larger, more irregular area in central, eastern and southern Brazil. Both areas have some elevations above 1000 meters. (Figure 1) An examination of the geologic history of these areas is presented with the hope of determining some constraints on models explaining anomalous topography.

Methods of Investigation

In this I have attempted to determine the timing and cause of uplift of the two major anomalous areas. Generalized (and when necessary, detailed) stratigraphy of the uplifted areas and their surrounding regions was examined to determine the timing of uplift. Depositional hiatuses or erosional surfaces in the highland stratigraphy, synchronous with increased deposition of highland derived clastics in the surrounding basins should provide constraints on timing of uplift. Progradation of river, with a dramatic increase in clastic input should provide evidence of accelerating uplift of their source areas. Therefore, rivers draining these areas were examined to see if seaward surges could be identified.

Structural features in the anomalous areas have been examined to discover possible reasons for uplift. Structures could provide pertinent information on stress regimes active prior to or during uplift.

GUYANA HIGHLANDS

Geologic Summary

The Guyana Highlands consist of a sequence of clastic rocks overlying a basement complex of metamorphic rocks. This basement complex is comprised of two units. The lower of these two units, the Imataca complex of Venezuela and the lower basement complex of British Guiana, is characterized by granitic and amphibolitic gneisses, iron bearing quartzites, and sericites, chlorite and talc schists and phyllites, pegmatites, gabbros, dolerites and spillites. (Lopez, 1956; Bracewell, 1956; Hurley and Rand, 1973) These rocks are early Precambrian in age, dated as 2700-3400m.y. by Rb-Sr whole rock analysis. (Hurley and Rand, 1973)

Overlying this complex is the Pastora series, consisting of volcanic rocks and volcanigenic and pelitic sediments of late Precambrian age, metamorphosed during the Trans-Amazonian Orogeny, approximately 1800-2100 m.y.b.p. (Hurley and Rand, 1973)

Unconformably overlying this basement system is the relatively flat lying, detrital, Roraima sequence. This consists of about 2400 meters of thinly and thickly bedded pink, yellow and white sandstones, reddish quartzitic sandstones and quartzites, red, green and white (porcellanite) jasper, shale, conglomerate and boulder conglomerate. The basal unit of the sequence is a ten meter thick diamond and gold bearing conglomerate. (Lopez, Bracewell, 1956) The Roraima is divided into upper, middle and lower units by red jasper horizons which appear conformable but are likely to disguise major depositional hiatuses. (Gansser, 1974) There Both the Proterozoic and Permo-Triassic dikes intrude the lower and middle Roraima members. The upper unit however, is unaffected. This suggests a post-Permo-Triassic age for the upper unit. Tentative fossil evidence, and stratigraphic correlation with sediments of the table mountains to the west suggest a cretaceous age (80m.y.) for the upper Roraima sediments.

(Gansser, 1974)

The Roraima sequence is found overlying much of the basement complex in eastern Venezuela and western British Guiana, and in general correlates with areas of high topography. The highest topography occurs in the Pakaraima Mountains which traverse the border, with Mt. Roraima as the highest peak. (Figure 2)

Basin Analysis Around the Guyana Highlands

As deposition of the Roraima formation occurred at least until the upper cretaceous, it is assumed that the time of uplift of the anomalous area is post-cretaceous. Therefore, an examination of the tertiary stratigraphy in the surrounding depositional basins was conducted in search of resedimented Roraima.

The eastern Venezuelan basin is located between the guyana shield on the south and the coastal ranges on the north. More than 12,000 meters of mesozoic and cenozoic sandstones, limestones and shales are preserved. (Hedberg, 1956; Liddle, 1946) These appear to be typical shelf sediments, with continental affinities increasing toward the guyana shield. Notable in this sequence is the Santa Ines formation of Miocene age. It is much thicker than any of the other cenozoic formations (3500-7000 meters thick). Overlapping beds and depositional inclination suggest rapid deposition. The unit is comprised of coarse grained sandstones and massive conglomerates containing chert, quartzite and limestone pebbles, and schist fragments. It lies locally unconformably above Oligocene or Upper Eocene sandstones and shales. (Liddle, 1946)

These miocene sediments appear to be thick fanglomerate deposits adjacent to a rapidly rising highland. Whether or not this highland was the guyana is not clear because the uplift of the coast ranges of Venezuela culminated in the Miocene. (Beck, 1978; Maresch, 1974) Resolution of this question will have to await dating of the metamorphic fragments, which

should be Precambrian if guyana is the source and cretaceous if the source is the coast ranges.

Another area containing possible resedimented highland strata is the northern coastal region of British Guiana. This area contains the tertiary white sand series. This is comprised of red, yellow and white fine grained sands, quartz gravels and reddish clays. These deposits attain a maximum known thickness of 2100 meters at the mouth of the Berbice River. The coastal deposits become restricted in area further inland. They appear to occupy the valleys and estuaries of an old river system. These sediments become coarser in proximity to the Pakaraima Mountains.

An important feature of the white sand coastal and alluvial deposits is their alluvial diamond content. (Bracewell, 1956) The only recognized source of these diamonds is the basal Roraima unit. (Lopez, 1956; Bracewell, 1956) As the headwaters of the tertiary river system appear to be in the Pakariama Mountains, it is likely that the white sand formation consists to a great degree of resedimented Roraima.

The age of the white sand deposits is possibly miocene-pliocene, as dated by foraminifera in arenaceous sediments drilled from a test well near New Amsterdam. (Bracewell, 1956)

An examination of the Orinoco and Amazon River deltas for indications of uplift of the Guyana Highlands did not reveal any correlations. The generalized survey of the tertiary of the Amazon basin (Bigarella, 1973) and the more detailed stratigraphic analysis of the delta area (Oliviera, 1956) do not indicate either the presence of resedimented highland clastics or massive progradation of the delta due to uplift of the Guyana Highlands.

Brazilian Highlands

Geologic Summary

The geology of eastern Brazil is comprised of outcropping Brazilian shield and the sediments of a variety of Phanerozoic basins. The

depositional basins are divided into six major stratigraphic sequences separated by regional scale unconformities. (Campos et.al., 1976) The rocks of the Brazilian shield are divided into lower, middle and upper Precambrian.

The lower Precambrian consists of metamorphosed granite rocks, pegmatites and dikes older than 2200m.y.b.p. and includes the Trans-Amazonian orogenic cycle. These rocks range from ultrabasic to granitic, including pegmatites and volcanics. The upper Precambrian consists of the Uracuano cycle (900-1400m.y.b.p.) and the Brazilian cycle (700-450m.y.b.p.) The Brazilian cycle involved intense orogeny, migmatization and intrusive activity. (Almeida et.al., 1973; Oliviera, 1956)

Campos et.al. (1974) distinguish a metasediment platform cover from the basement complex. This consists of early to late Cambrian arkoses, pelites and limestones, partly deformed during the last episodes of the Brazilian Orogeny, and Cambro-Ordovician silicic igneous rocks.

The six phanerozoic depositional sequences are found in both the intercratonic and coastal basins. However, the lower three sequences are much better developed in the intercratonic basins than in the coastal basins. The first sequence is comprised of thick coarse sandstones and basal conglomerates overlying the eroded metamorphic surface. These sediments were deposited in the topographic depressions resultant from the Brazilian foldbelts, and range in age from Silurian through Devonian.

Sequence II is deposited over a glacial erosion surface developed on sequence I, and continued basin filling from lower Carboniferous to Permo-Triassic. Cycles of sandstones, evaporites, carbonates and black shales with the interior basins suggest fluctuations of climate from arid to humid.

Sequence III is separated from the underlying sediments by a major erosional unconformity which locally stripped sedimentary cover to the basement. The sequence contains pre-rift red beds (shales and sandstones)

and rift filling fluviclacustrine sediments. It is of upper Jurassic to lower Cretaceous age. Thick igneous basalt flows and diabase dikes and sills from 110-148m.y. old, associated with syntectonic sediments are found in the central and southern basins.

Sequence IV ranges from Aptian to Santonian age and contains localized evaporites and limestones in the interior basins and thick evaporitic sections in the coastal basins. These are underlain by continental basal conglomerates and sandstones. The evaporites grade upward into beach sandstones, limestones and marine shales. This is the first recorded marine sedimentation on the present Atlantic margin of Brazil, however, the equatorial coastal basins contain no evaporites from this sequence. (Campos et.al., 1974)

Sequences V and VI are not distinguished in sediment type and are only locally separated by an Oligocene-Miocene unconformity. Throughout the cenozoic, terrigenous sandstores and shales, with some limestones, are common. Beach, shelf and slope facies of a prograding continental margin, undergoing oscillatory transgressions and regressions are represented by these sediments. (Ponte and Asmus, 1976)

Relation of Topography to Structures

Ponte and Asmus (1976) describe three general categories of elevated topography: tablelands (tabulieros and chapadas), cuestas and ridges. Tableland are sedimentary plateaus formed by flat lying Mesozoic and Cenozoic strata. These are presently bounded by erosional cliffs. Cuestas are open basins with shallow inward centripetal dips. Ridges are elongate areas of high, rugged topography. These features are present in figure 4a. The authors suggest that many of these topographic features are bounded by or localized on reactivated basement structures.

The Tucano, Jatoba and Araripe tablelands, about 500-700 meters high, have some boundaries which coincide with Precambrian lineaments in the

basement floor. (Figure 4b) (Ponte and Asmus, 1976)

The uplift of the Serra do Mar region began in the upper Cretaceous along preexisting Cambro-Ordovician (Brazilica Orogeny) fault trends. It is likely that these faults resulted from the reactivation of Trans-Amazonian structures. Two other tectonic episodes occurred during the Oligocene and Pliocene, accentuating relief and initiating intensive erosion. Current seismicity in the Serra do Mar indicate that the region is still somewhat active. (Almieda, 1976)

Although it is clear that basement trends have some influence on the form of the small scale uplifts mentioned, other areas of high topography are unaffected. It is likely that basement reactivation is only a response to some other mechanism of uplift.

Basin Analysis Around the Brazilian Highlands

An examination of the Tertiary sediments in the coastal basins of Brazil reveals a thick continental clastic section deposited in Miocene-Pliocene times. (Figure 5) The Barreiras is well developed in the Reconcavo, Sergipe-Alagoas, Potiguar and Recife-João-Pessoa basins. (See Figure for basin locations.) In the Recife and Reconcavo basins, the clastics overlie Miocene carbonates and shales. In the Sergipe-Alagoas basin, Miocene-Oligocene marine sediments are thin or absent. In the Potiguar basin, the Barreiras group unconformably overlies Maestrichtian carbonates. (Asmus and Ponte, 1973)

Almeida (1976) correlates the Barreiras group with continental Pliocene deposits in the Santos basin. He suggests that the areal extent of these continental clastics is indicative of uplift of all of central and eastern Brazil in the Pliocene.

Judging from the data available it is likely that much of the anomalously high topography was uplifted in Pliocene times. Pre-Pliocene uplifts probably also occurred, although perhaps not on such a regional scale.

Igneous Activity

Southeastern Brazil is an area of extensive alkaline and basaltic magmatism. (Figure 6a) A period of early cretaceous basaltic extrusion and diabase dike and sill injection is associated with the opening of the South Atlantic. Northeastern Brazil was unaffected by this volcanism. (Campos et.al,)

A second, late cretaceous to early tertiary igneous activity, although less widespread, resulted in thick volcanic accumulations in central and southern Brazil. This younger volcanism has been associated with igneous activity along various oceanic fracture zones. (Figure 6b) (Ponte and Asmus, 1976)

Conclusions

Stratigraphical analyses of the depositional basins surrounding the two anomalous areas suggest that the Guyana Highlands were uplifted between late Cretaceous and Miocene times and the Brazilian Highlands were uplifted in pre-Pliocene (probably Miocene) times. Although the timing of uplift is fairly well constrained, the causes and mechanisms of uplift are much less certain.

Two hypotheses have been submitted by Burke et.al. (1977): uplifting in response to a mantle hot spot and uplift due to intraplate stresses.

A hotspot mechanism for the uplift of the Brazilian Highlands may have some validity. The peak of magmatic activity at about 65m.y. could be associated with the initial stages of uplift which is still in effect today. Erosion would level any high topography in a few million years which was not constantly and dynamically maintained. If volcanism is an accurate indicator of high heat flow in the lithosphere, then the sporadic igneous activity in the Brazilian Highlands does not provide compelling evidence of hotspot activity continuous from 80m.y. to present.

The Guyana Highlands contain no cenozoic magmatic rocks. Thus a

hotspot mechanism for this uplift is unlikely.

Uplift as a response to intraplate stress is difficult to quantitatively constrain because the state of stress in the lithosphere is unknown. Examination of anomalously high topographic areas on both sides of the Atlantic suggest that crude correlations can be drawn between the age of the terrains being uplifted and their form. Areas of old Precambrian crystalline rocks, such as the Brazilian, Guyana, and Canadian Shields seem to be characterized by broad uplifts. Paleozoic and younger mountain belts tend to form narrow elonate uplifts correlative with the shape of the Orogen. If both types of areas are responding to the same intraplate stresses, the difference in their form may be the result of crustal anisotropies. To a first approximation, the shield areas respond in a homogeneous fashion while the younger mountain belts behave anisotropically. The old structures preserved in the shield areas seem to have second order effects on the boundaries of the elevated areas in Brazil, but little evidence is found for this in Venezuela.

As the two anomalous areas are large in scale, it is unlikely that a model explaining their origin can be formed from an examination of South America alone. Any model proposed for the formation of these broad topographic uplifts on the continents must result from a worldwide analysis.

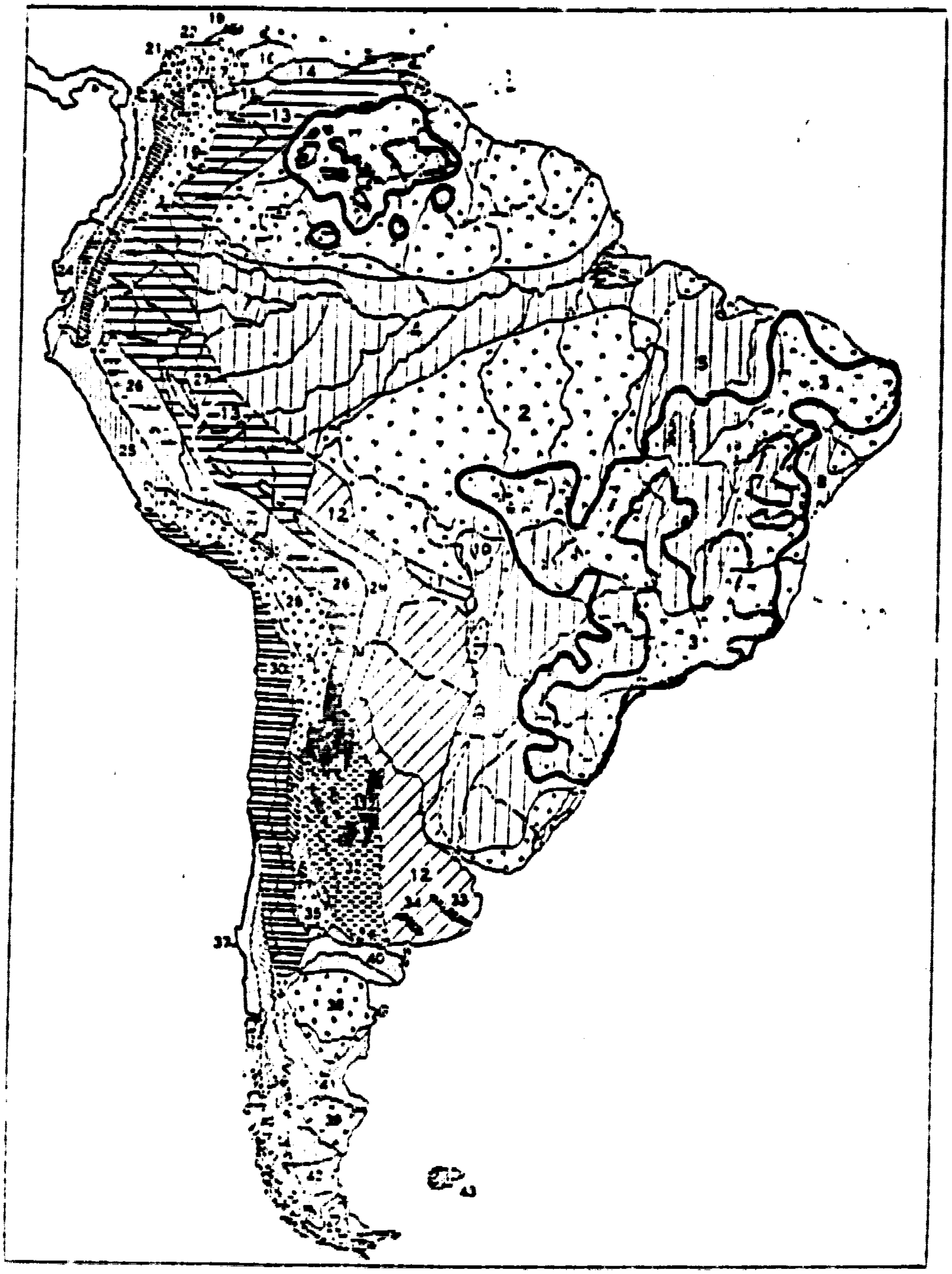

Figure 1

Location of anomalous topographic areas of South America. a) Guyana Highlands; b) Brazilian Highlands.

> 500 meters

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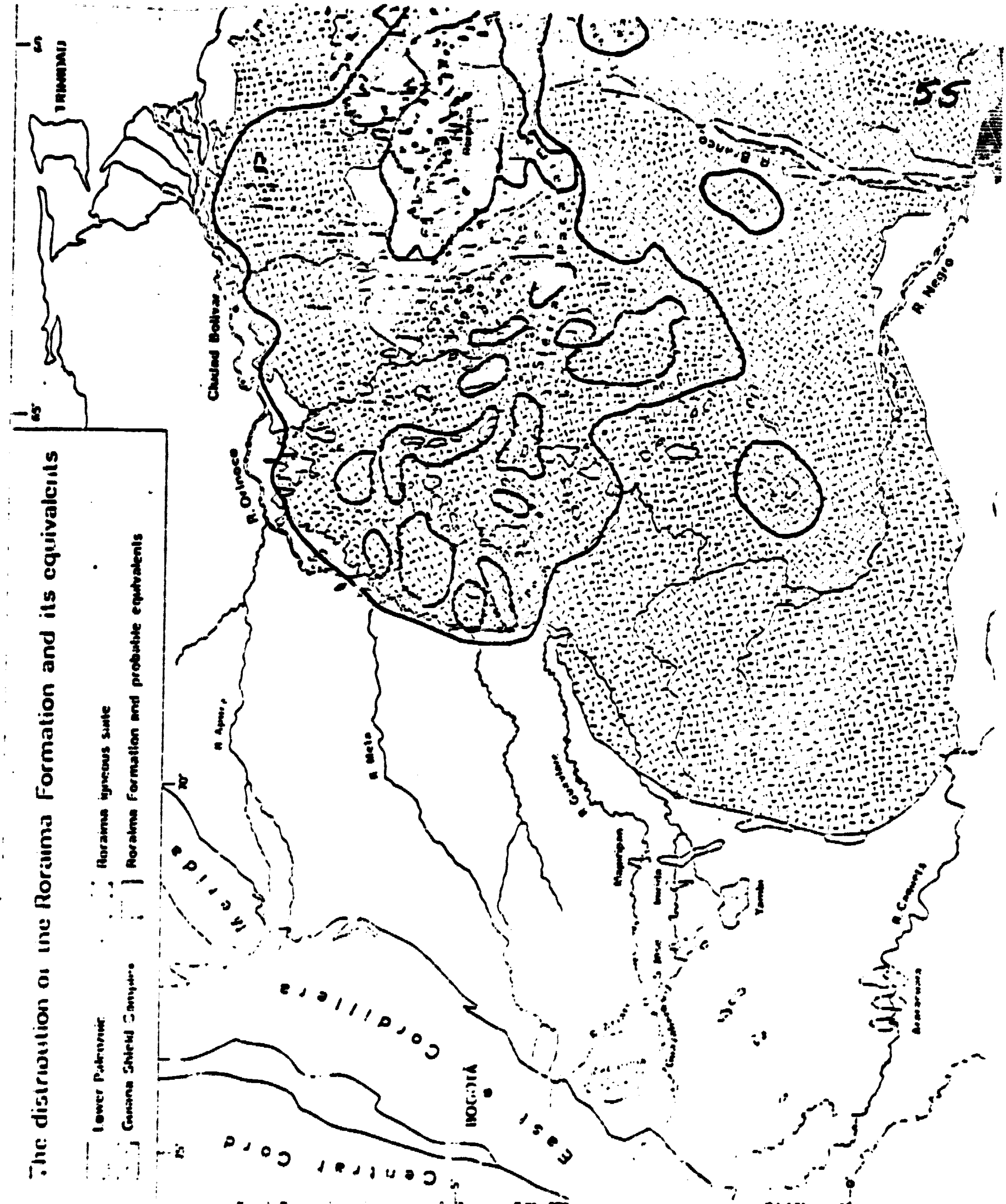

Figure 2

Distribution of Roraima formation versus high topography.

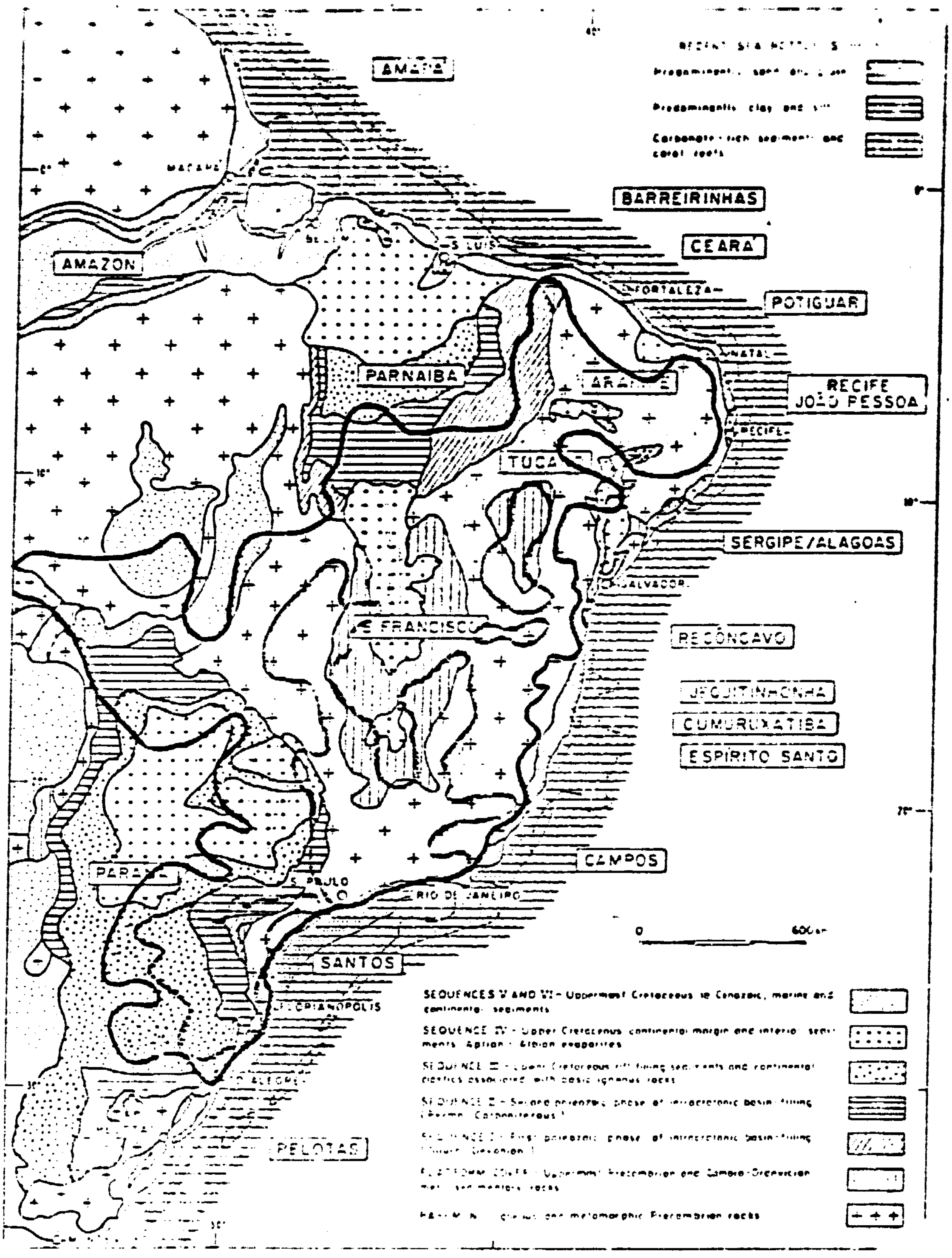

Figure 3

Basin locations, geologic summary and high topography of the Brazilian Highlands

[ ] > 500 meters [ ] > 1000 meters

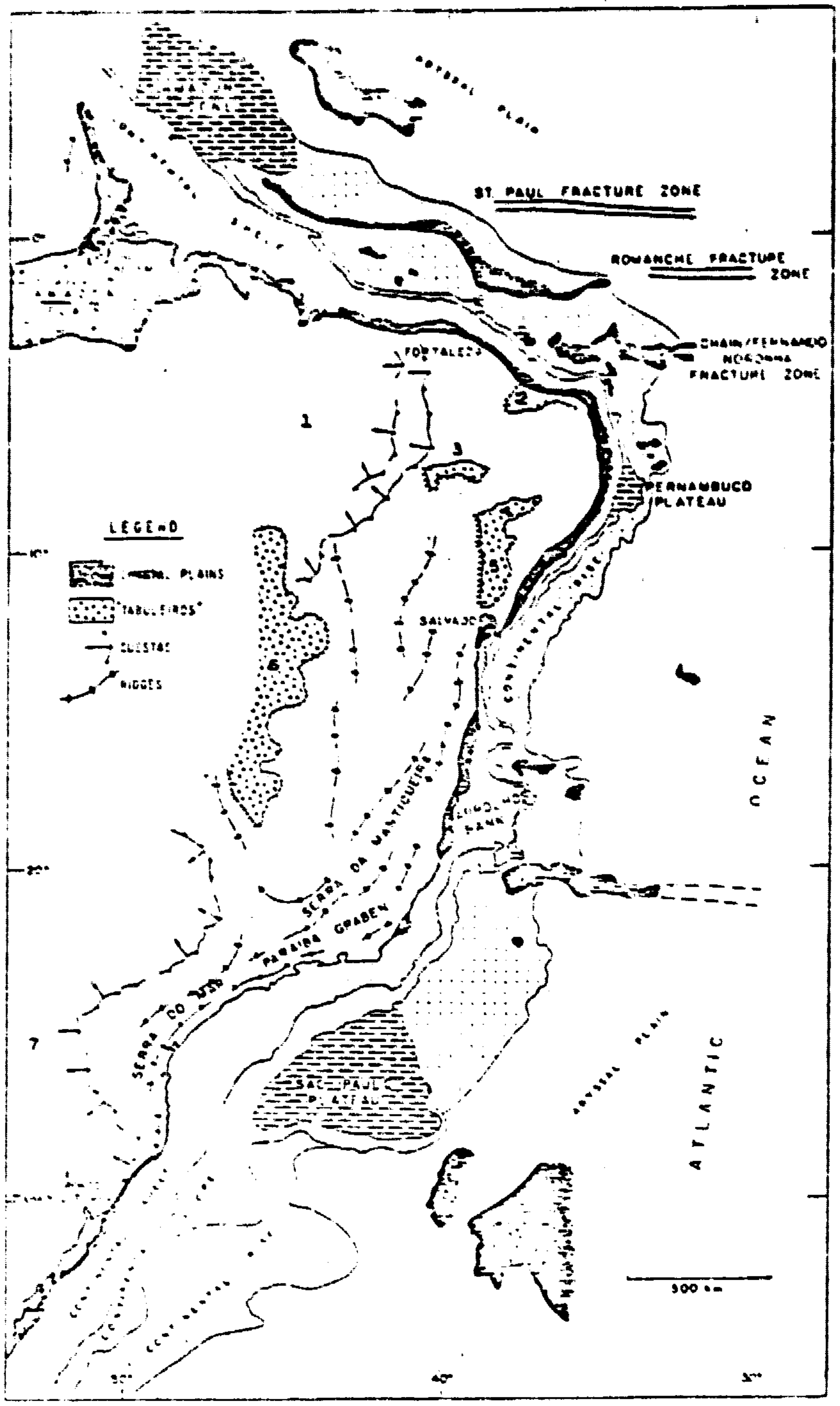

Fig. 6 — Physiographic map of the Brazilian continental margin. 1 — Maranhão sedimentary basin; 2 — "Chapada" Apodi; 3 — "Chapada" Araripe; 4 — "Tabuleiros" Jatobá; 5 — "Tabuleiros" Tucano; 6 — "Chapadões" Urucuia, 7 — Paraná basin. (Modified from Asmus, 1973).

Figure 4a

Location of tablelands, cuestas and ridges.

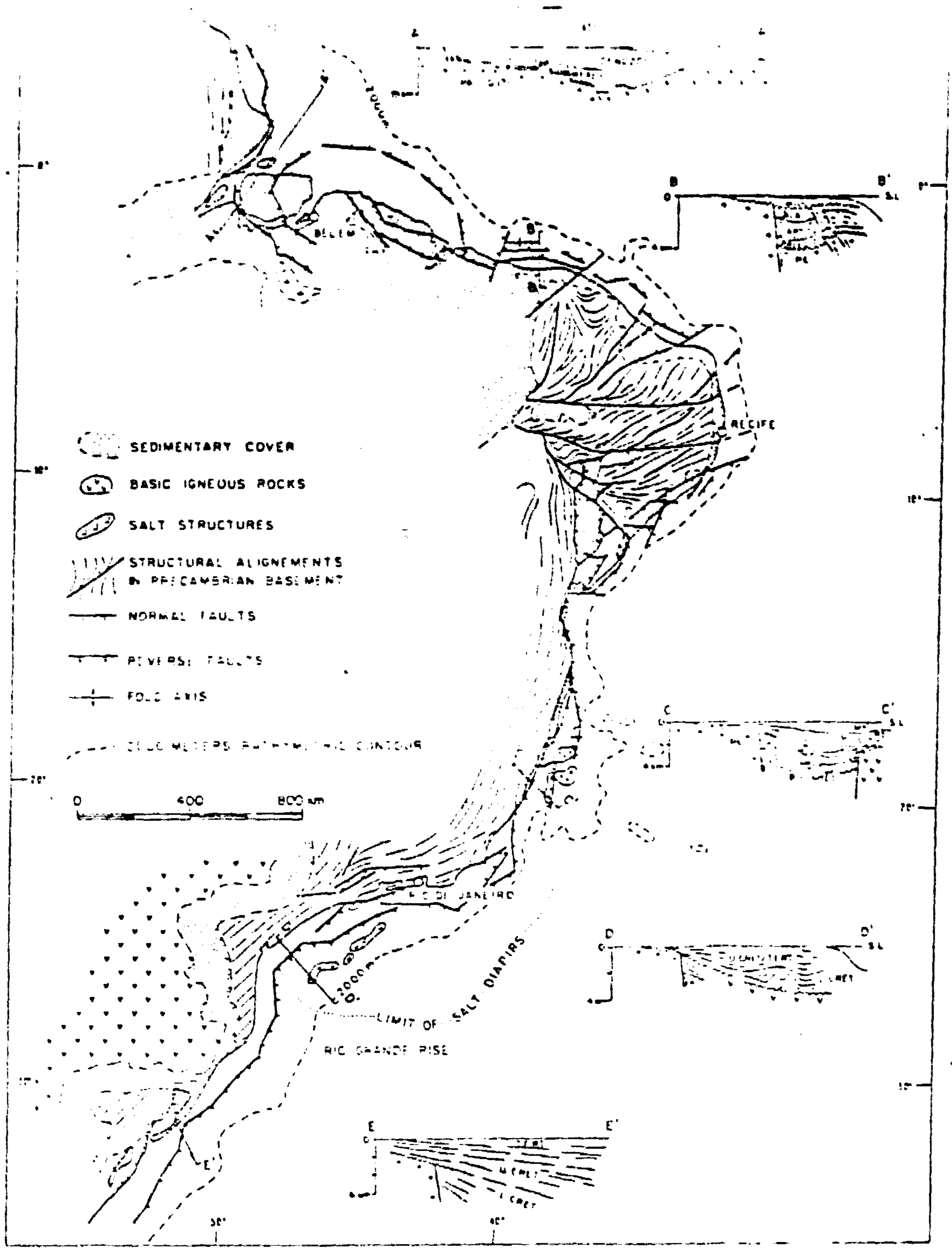

Fig. 4. Structural framework of the Brazilian Shield and continental shelf. Cross-sections show the general profiles of some typical coastal basins.

Figure 4b

Precambrian basement structures.

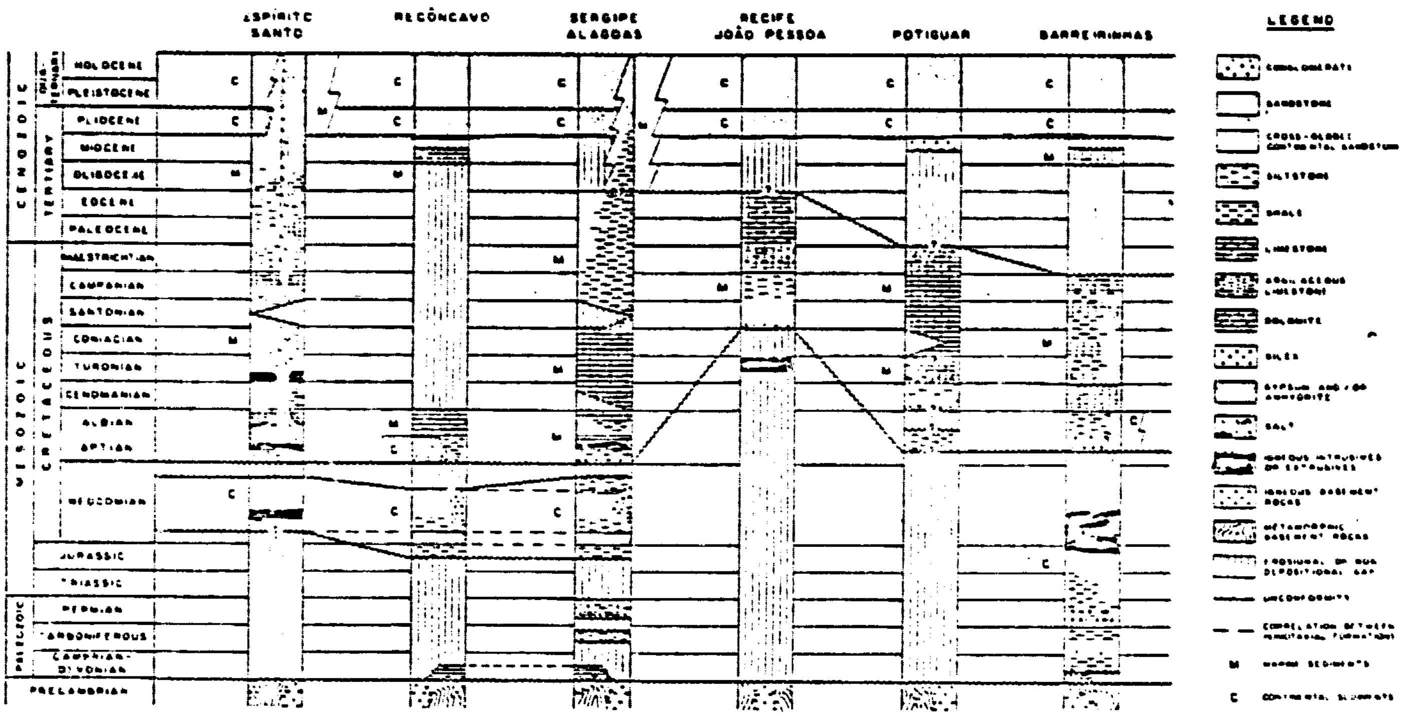

Fig. 2. Stratigraphic correlation chart from Barreirinhas basin to Espirito Santo basin.

Figure 5

Base of Barreiras formation.

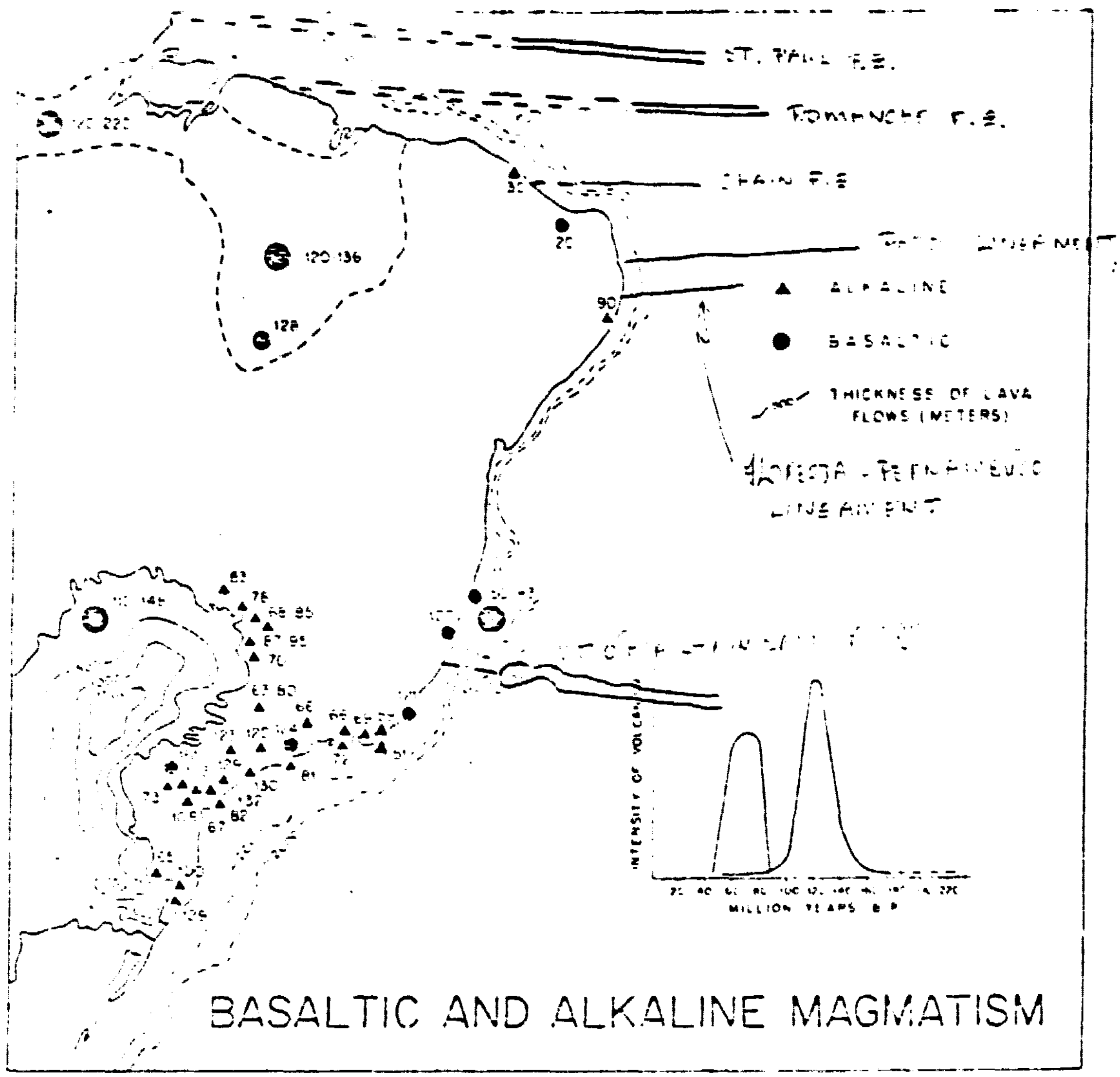

Fig. 5. Geographic and age distribution of basaltic and alkaline magmatism in Brazil. Two prominent peaks of volcanic activity are observed. Early Cretaceous (110-140 my) and Late Cretaceous to Early Tertiary (50-80 my) (after Asmus, 1973).

Figure 6a

Distribution of magmatism and fracture zone trends.

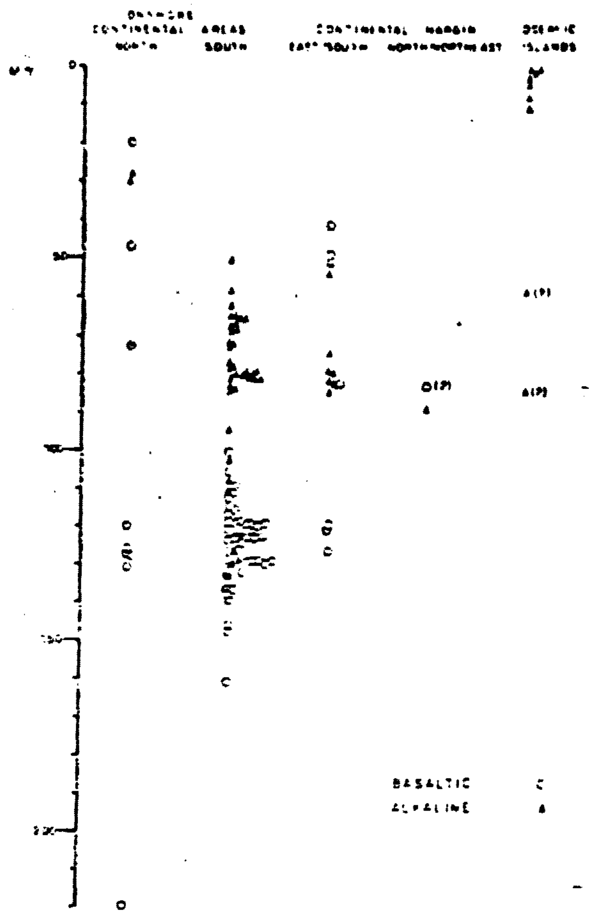

Fig. 10 — Age and regional distribution of the Mesozoic-Cenozoic volcanism in Brazil. Data from Hennies and Hasui (1968); Cordani and Hasui (1968); Cordani and Blazekovic (1970); Hasui and others (1971); Damasceno (1966); Cordani (1970); Amaral and others (1967); Vandoros and others (1966); Ribeiro and Cordani (1966). (After Asmus, 1973).

Figure 6b

REFERENCES

de Almeida, F.F.M., 1976. The system of continental rifts bordering the Santos Basin, Brazil: continental margins of Atlantic type, F.F.M. deAlmeida, ed. An ais da Academia Brasileira de Ciencias, vol. 48, Sao Paulo.

de Almeida, F.F.M., Amaral, G., Cordani, U.G. and Kawashita, K., 1973. The Precambrian evolution of the South American cratonic margin south of the Amazon River: the ocean basins and margins, vol. 1, The South Atlantic, A.E.M. Nairn and F.G. Stehli, eds., Plenum Press, New York.

Asmus, H.E. and Ponte, F.C., 1973. The Brazilian marginal basins: the ocean basins and margins, vol. 1 The South Atlantic. Plenum Press, New York.

Beck, C.M., 1978. Polyphase Tertiary tectonics of the interior range in the central part of the western Caribbean chain, Guarico State, northern Venezuela. Geologie en Mijnbouw, vol. 57.

Bigarella, J.J., 1973. Geology of the Amazon and Parnaiba Basins; The Ocean Basins and Margins, vol. 1, The South Atlantic. A.E.M. Nairn and F.G. Stehli, eds., Plenum Press, New York.

Bracewell, S., 1956. The Guianas-British Guiana: Handbook of South American Geology, W.F. Jenks, ed. Geological Society of America, New York.

Burke, DeLong, Dewey, and Kidd, 1977. Global Tectonic Studies.

Campos, C.W.M., Ponte, F.C. and Miura, K., 1974. Geology of the Brazilian Continental Margin: the Geology of the Continental Margins, C.A. Burk and C.L. Drake, eds., Springer-Verlag, New York.

Frick, U. and Steiger, R.H., 1974. K-Ar dating of some Guyana shield dyke rocks, Verhandl. Naturf. Ges. Basel, vol. 84, Basle.

Gansser, A., 1974. The Roraima Problem (South America), Verhandl. Naturf. Ges. Basel, vol. 84, Basle.

Hedberg, H.D., 1956. Venezuela-northeastern Venezuela: Handbook of South American Geology, W.F. Jenks, ed. Geological Society of America, New York.

Hurley, P.M. and Rand, J.R., 1973. Outline of Precambrian Chronology in Lands Bordering the South Atlantic, Exclusive of Brazil: The Ocean Basins and Margins, vol. 1, The South Atlantic, A.E.M. Nairn and F.G. Stehli, eds., Plenum Press, New York.

Liddle, R.A., 1946. The Geology of Venezuela and Trinidad. Cayuga Press, New York.

Lopez, V.M., 1956. Venezuela-Venezuelan Guiana: Handbook of South American Geology, W.F. Jenks, ed. Geological Society of America, New York.

Maresch. W.V., 1974. Plate tectonics origin of the Caribbean mountain system of northern South America. GSA Bulletin, vol. 85.

de Oliveira, A.I., 1956. Brazil: Handbook of South American Geology, W.F. Jenks, ed., Geological Society of America, New York.

Ponte, F.C., and Asmus, H.E., 1976. The Brazilian marginal basins: current state of knowledge: continental margins of Atlantic type, F.F.M. deAlmeida, ed. Anais da Academia Brasileira de Ciencias, vol. 48, Sao Paulo.

Enclosure 4 64

EOS vol. 60, no. 6, February 6, 1979, pages 59-64

Plateau Uplifts: Mode and Mechanism

This meeting report was prepared by T. R. McGetchin (Lunar and Planetary Institute, Houston, Texas), K. Burke (State University of New York, Albany), G. Thompson (Stanford University, Stanford, California), and R. Young (State University of New York, Geneseo)

Introduction

One of the keys to understanding the dynamics of plate interiors is the puzzle of plateau uplifts, especially those isolated from active mountain belts. These are of particular interest because they apparently involve deeply rooted processes and are independent of active subduction. Some of these areas are capped by young alkaline volcanics, while some are free of volcanics. Putorana in Siberia, the Adirondacks and Black Hills of North America, and the Serra do Mar in Brazil are typical cases—irregular areas 100-200 km in diameter, 1 km above their surroundings. Volcanic-capped uplifts are common in Africa (Ahaggar, Tibesti, Jos Plateau, Ngaoundéré, and the Cameroon Zone), as are examples of the volcanic-free uplifts (Fouta Djallon, Angola.

TABLE 1. Examples of the Types of Tectonic Settings of Plateaus of the World

Plate Margin: Convergent	Plate Margin: Divergent	Intraplate
	Continental Plateaus	
Tibet	Ethiopia	Colorado Plateau
Iran		
Shillong		
Altiplano		
	Oceanic Plateaus	
Caribbean	Iceland	Hawaii
	Galápagos	

Adamawa, and high Veldt). The Colorado Plateau may be similar to volcanic-capped African uplifts.

In order to address the origin of these intriguing features, a conference on Plateau Uplifts: Mode and Mechanism was held August 14–16, 1978, in Flagstaff, Arizona, cosponsored by Working Group 7 of the Inter-Union Commission on Geodynamics and the Lunar and Planetary Institute. The meeting was hosted by the U.S. Geological Survey. Sixty participants spent three days discussing various aspects of plateau uplifts: geomorphology, structural geology and tectonics, deep crustal and upper mantle structure from geophysics, petrologic and geochemical constraints on models, and models for plateau uplifts. While discussion naturally focused on the Colorado Plateau, uplifts on virtually all the continents were discussed as were plateau features on the sea floor and the planet Mars. About 50 of the participants enjoyed a 2-day premeeting field trip from Las Vegas, Nevada, to Flagstaff, Arizona. Highlighted were structural and stratigraphic relations along the southwestern edge of the Colorado Plateau viewed by boat in the Lake Mead region.

Uplift Mechanisms

More than a dozen different mechanisms were proposed to account for the plateau uplifts during this conference, which in some form or other could be applied elsewhere in the world. These include (1) thermal expansion of the lithosphere due to a deep mantle plume or hot spot, (2) thermal expansion due to overriding and subduction of a ridge, (3) thermal expansion due to shear heating, accompanying relative motion along the lithosphere-asthenosphere interface, (4) volumetric expansion accompanying partial melting, (5) hydration reactions within the mantle, such as serpentinization, (6) introduction of volatiles from a deep-seated source beneath, due to heating or some other dehydration mechanism affecting high-pressure hydrous phases such as phlogopite, amphibole, or humite, (7) depletion of 'fertile mantle' in garnet and iron due to partial melting, producing a chemically depleted, refractory residuum of lower density than the original garnet lherzolite, (8) 'crustal thickening' due to a horizontal mass transfer of material (by an unspecified process), (9) underplating or subduction at a very shallow (or even horizontal) angle, (10) metamorphic (solid state) reactions such as basalt-eclogite or spinel-olivine, (11) 'simple' subduction, (12) cessation of subduction and complex reactions accompanying thermal reequilibration of a slab which has stopped, (13) reactivation of preexisting low-angle (listric) thrust faults,

TABLE 2. Major Plateaus and High Plains of the World (Kossinna, 1933)

	Mean Elevation, m	Area, 1000 km²
German Subalpine Foreland	500	35
Iceland Plateau	600	70
New Castille Plateau	600	60
Old Castille Plateau	700	70
French Massif Central	700	70
Scandinavian Highlands	700	350
Deccan Plateau	800	400
Shotts (Atlas) Plateau	800	80
Nejd Plateau (Arabia)	900	700
Anatolian Plateau (Asia Minor)	1,000	500
Kalahari	1,000	2,100
Tarim Basin	1,100	600
Gobi	1,100	1,650
East African Lakes Plateau	1,200	1,000
Iranian Plateau	1,300	1,000
Great Basin (United States)	1,500	600
Colorado Plateau	1,800	500
Greenland Ice Plateau	1,900	1,870
Armenian Highlands	2,000	300
Yunnan Highlands	2,000	300
Mexican Plateau (Altiplano)	2,000	350
Ethiopian Plateau	2,200	450
Antarctic Ice Plateau	2,500	12,800
Ecuador Plateau (Altiplano)	3,000	15
Bolivian Plateau (Altiplano)	3,800	350
Pamir Plateau	4,000	100
Tibetan Plateau	4,500	2,000
Tharsis area on Mars	10,000	16,000

and finally (14) detachment and foundering of large pieces of the lithosphere beneath the continents (termed 'delamination') and flow of mantle material in the asthenosphere to replace it. (The large subterranean frog of classic Chinese and Japanese literature whose burping was believed to be responsible for earthquakes and volcanic eruptions went unnoticed; an oversight by the program committee.) The consensus seemed to be that on the continents, uplifts of regional extent are invariably related to plate motions—subduction or fission. On the ocean floor they are not. On Mars we have insufficient data to discuss the problem intelligently, tempting though it is to speculate on the possible implications of Tharsis for Martian plates in some incipient form.

Uplifts of the World

Since the term 'uplift' implies upward structural displacement, there is commonly ambiguity and argument regarding what constitutes an uplift, in contrast to an area which is topographically high for a variety of other reasons. This is particularly true in areas where geological data are sparse. Following Holmes, plateaus are defined as broad uplands of considerable elevation. Terrestrial plateaus occur both on the continents and on the ocean floor. Some are associated with plate margins both convergent and divergent, and others are not, as Table 1 shows.

Broad areas of sea floor stand significantly above the mean depth; these are of considerable interest because the sea floor, in some sense at least, is simpler than the continents. Crough developed a theory of ocean rises illustrated by the Hawaiian Swell. The lithosphere thins abruptly as it moves over hot spots and then thickens by slow cooling as it moves away, controlled by the same processes as lithosphere moving away from a spreading ridge. The rapid heating cannot be accomplished by conduction but may be caused by intrusion of dikes into the lithosphere.

Plateaus of the sea floor may owe their elevation to any of several causes: some are volcanic piles (Manihiki and Ontong Java plateaus in the Pacific), some are believed to be continental fragments (Seychelles), and others, thermal perturbations (expanded mantle). Problems occur when these features encounter active subduction zones. The floor of the Caribbean is believed to be occupied by such a buoyant residuum of ocean floor plateau.

Continental uplifts (Table 2), because of intense erosion rates, are by definition young, generally less than about 40 m.y. old. (Hence the oldest plateaus on earth are to be found on the sea floor—at substantial distances from the active spreading centers). Continental plateaus are commonly, but not exclusively, associated with plate margins. An example of a plateau at divergent plate boundaries is in Ethiopia; those on convergent plate boundaries are in Tibet, Iran, and the Shillong and Altiplano plateaus. The latter are associated with active convergence.

Continental uplifts also exist, however, which are not associated with plate margins, namely, those lying within continents and remote from plate boundaries. A number of these exist in the world, some of which were discussed and summarized.

An interesting and important insight on the question of the relative uplifts of large continental areas is provided by comparing data from the stratigraphic record on the percentage of continental areas flooded as a function of time and by use of hypsometric curves to infer the corresponding uplift. Use of this method (Bond) suggests extensive post-Miocene uplift of Africa, somewhat older (pre-Eocene) uplift of western North America, extensive uplift of Australia in Cretaceous time which had subsided substantially by Eocene time, and post-Cretaceous uplift of central Europe. The timing of such intraplate plateau uplifts in relation to other regional and global tectonic events is important. For example Burke pointed out that the great African plateaus appear to begin in the early Tertiary, a time when sea floor data elsewhere suggest that the Africa plate had come to rest with respect to the underlying convection pattern.

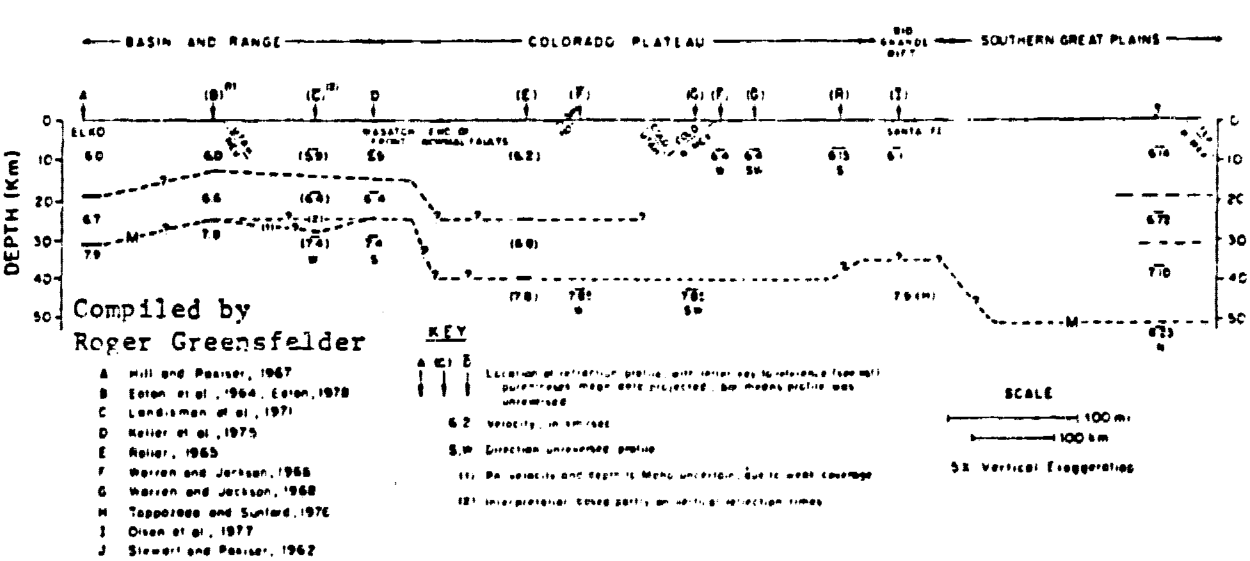

Fig. 2a. Geophysical properties along an east-west cross section through the Colorado Plateau. The crustal thickness beneath the Colorado Plateau is greater than that on either side; its heat flow is above normal; its electrical conductivity high, and the p wave velocity is low. (Reprinted from *Papers Presented to the Conference on Plateau Uplift: Mode and Mechanism*, pp. 52-54, Lunar and Planetary Institute, Houston, Texas (Thompson and Zoback, 1978).)

Several other plateaus within continental areas of the world were discussed at the conference. The high plateau of central Mexico is certainly related to the rest of the American Cordillera. Fucugancni (abstract only) suggests that the Mexican volcanic belt is the result of the subduction of the Cocos plate at the mid-America trench in middle to late Tertiary time and suggests that subduction of the East Pacific Rise may occur in the near future. In this sense the Mexican highlands may be an important clue to the mid-Tertiary (pre-30 m.y.) history of the western United States.

The Rhenish shield, in western Germany and Luxemburg, is a relatively small (200 × 100 km) but interesting area because it has undergone some 150 m of uplift during the Quaternary. It is bounded on all sides by seismically active rift valleys and by young volcanism (Eifel area). An ambitious multidisciplinary investigation is underway; the current working model (Illies) suggests that the Rhenish shield is a fault-bounded structural block produced by northward displacement of the Alps during the late Tertiary and that the uplift and volcanism are due to shear heating resulting from slip at the base of the lithosphere. A comparision of the Rhenish and Fennoscandian shields (Theilen and Meissner) shows that the Moho and asthenosphere are considerably deeper under Fennoscandia; it is suggested that Fennoscandia's uplift is due to its close association with the Atlantic, possibly involving mantle creep.

The Indian subcontinent (Kailasam) is flanked on the north by the Tibetan plateau, the largest and highest in the world—this is the convergent plate boundary between the continental plates of India and Eurasia; the Shan plateau to the east is smaller but also has experienced uplift over an extensive area. Four smaller but substanal areas within peninsular India have undergone epeirogenic uplift during the Cenozoic—the Deccan and Karnataka in south India and Chota Nagpur and Shillong plateaus in eastern and northeastern India. All are being investigated for both scientific and economic reasons; geodesy, seismicity, gravity, and volcanic geology suggest that these plateaus are active. A regional negative gravity anomaly is believed to indicate an unusually hot upper mantle and deeply rooted causes for the uplifts, such as plumes, hot spots, and thermal expansion.

The uplifts of central Asia (Baikal and Tien Shan) were described as a SW-NE striking arch. Rifts are associated with these uplifts, confined to their axes. Normal faults and earthquake local mechanisms suggest that the Baikal Rift is an extensional feature oriented across the arch. The upper mantle beneath the arch has low P_n, high Q, and low density, and it is in approximate isostatic balance. It is inferred that the asthenosphere extends to the base of the crust under the uplift. Lateral difference in lithosphere thickness alone, however, does not appear to account for the data, and laterally heterogeneous asthenospheric properties are implied, namely, a density contrast of 0.005 g/cm³. The Soviet investigator Zorin concludes that SW-NE compression (from Tibet) cannot generate all the structures but that uprise of hot mantle is required.

The Tharsis region on Mars is a large (16 million km²) area in which the structural uplift is believed to be about 10 km. A regional free air gravity anomaly of 500 mgal is associated with Tharsis. Its surface is covered by apparently young volcanoes and volcanic plains. Phillips described a number of possible evolutionary scenarios, but the important point is that plateau uplifts do occur in planets without well-developed plate tectonics.

Colorado Plateau

The Colorado Plateau occupies most of eastern Utah and parts of Colorado, Arizona, and New Mexico (Figure 1, cover illustration). Most rocks now exposed are undeformed Mesozoic sediments at an average elevation of about 2 km above sea level. Tertiary igneous rocks that include basaltic to silicic volcanics, kimberlite and lamprophyre dikes, and intermediate composition intrusive laccolithic centers are exposed within the province. Basaltic volcanism predominates and is associated with tensional structures around the plateau margins; the source of the basalts is mantle peridotites (Moore). Xenoliths and geophysical data together provide fairly specific constraints on the nature of the lower crust and upper mantle; the upper mantle is apparently hot, possibly even partially molten.

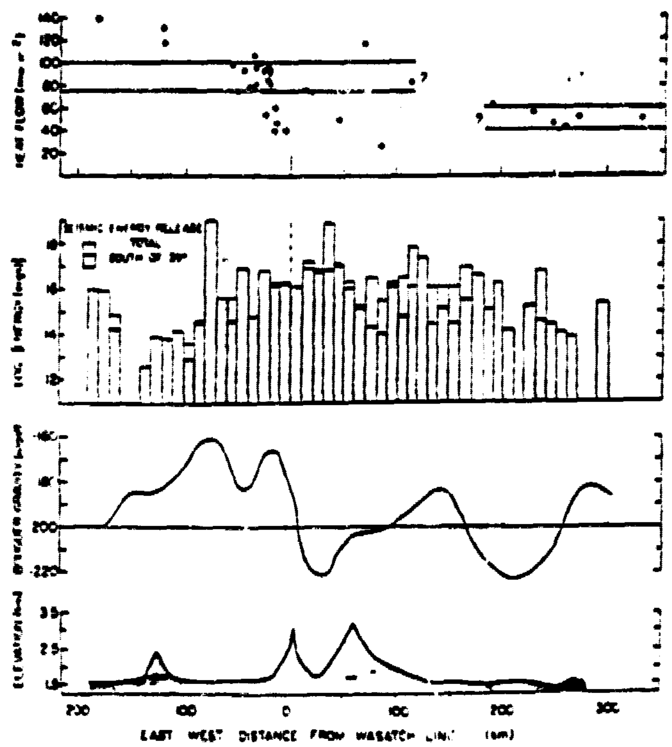

Fig. 2b. Summary of geophysical properties through the Colorado Plateau-Basin-Range province boundary (zero on abscissa) (Reprinted from *Papers Presented to the Conference on Plateau Uplift: Mode and Mechanism*, pp. 10–12, Lunar and Planetary Institute, Houston, Texas (Chapman et al., 1978)).

The lower crust and upper mantle xenoliths found in volcanic rocks show hydration effects (D. Smith) not observed around the plateau margins (Padovani, Hall, Simmons). Xenoliths suggest that the lower crust consists of high-rank mafic metaigneous rocks and the upper mantle of spinel peridotite at shallow depths and garnet peridotite at greater depth (Kay and Kay; Silver and McGetchin). An assortment of apparently deep-seated eclogitic rocks are present, exhibiting low-temperature, moderate-pressure mineralogy but with both

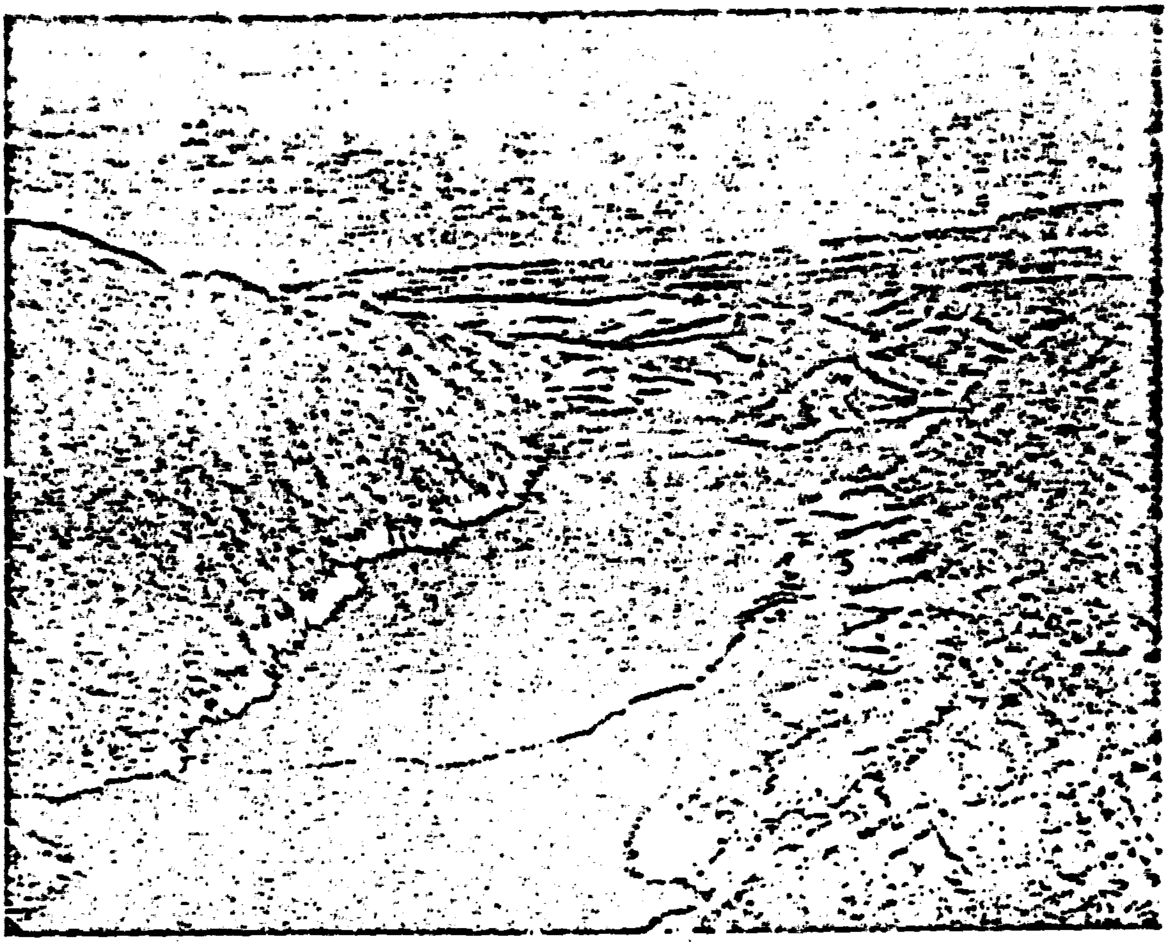

Fig. 3. Aerial photograph of the Lake Mead region showing the geological features of the transition between the Colorado Plateau (east, toward right) and the Basin-Range provinces (left). The view is northward in Iceberg Canyon, Lake Mead, showing tilted paleozoic and Precambrian rocks overlain by Muddy Creek Formation capped by lavas in Grand Wash. (Courtesy of U.S. Bureau of Reclamation.)

prograde and retrograde textures (Helmstaedt and Schulze), which may be fragments of a subducted plate.

New surface wave data (Keller, Braile, and Morgan) indicate that the crust under the Colorado Plateau is 45 km thick and is shieldlike, whereas the normal faulted regions surrounding the plateau have thinner crusts. A very low P_n velocity (7.5 km/s) within the transition zone was indicated by new refraction data. The upper mantle p wave velocity under the plateau is anomalously low, 7.8 km/s. A refraction line near Socorro in the Rio Grande Rift, just off the southeast margin of the plateau, revealed a crustal thickness of 33 km, an apparent P_n of 7.6 km/s, and a strong intracrustal reflection where a magma chamber had been postulated independently by Sanford and by deep seismic reflection (COCORP). New, deep-borehole heat flow measurements (Reiter, Mansure, and Shearer) reaffirm that the surface flux in the plateau is about 70 mW m^{-2}, or 1.6 HFU (μcal cm^{-2} s^{-1}). Local exceptions seem to be due to hydrothermal circulation and recent volcanism. This flux is higher than the average for areas of Precambrian crust and suggests that the crust and upper mantle beneath the plateau are unusually hot. As Thompson and Zoback showed, magnetic anomalies indicate a more conductive, presumably hotter, upper mantle beneath the plateau. The expansion caused by this heating may support the elevation of the plateau. The narrow zones of even thinner lithosphere which surround the plateau may supply the force necessary to cause the observed E-W compressive stress in the plateau and to keep the plateau from breaking apart. Chapman, Furlong, Smith, and Wechsler outlined the geophysical constraints on the nature of the boundary between the Colorado Plateau and Basin-Range (see Figure 2), namely, abrupt changes in seismic energy release, heat flow, crustal thickness, and P_n velocity but absence of significant elevation difference.

Cenozoic gravels and mid-Tertiary volcanics are crucial for interpreting the specifics of the uplift—its timing and location (Otton and Brooks, Young). In fact, there is some disagreement about whether there is significant relative uplift between the Colorado Plateau and Basin-Range provinces, both of which are structurally high (Damon; Pierce, Damon and Shafiqullah). There is no doubt that the entire region (western United States) is structurally as well as topographically elevated in relation to sea level or the midcontinent. A structure contour map drawn on the top of the Precambrian would show that the mountains of the Basin-Range stand even higher than the Colorado Plateau. Hence while the Colorado Plateau is a rather dramatic topographic high, its structural elevation relative to the Basin-Range province on the west, south, and east (Rio Grande Rift) is lower. The current view (Lucchitta; Shoemaker; Young) is that

EOS, vol 60, no 6, February 6 1979 page 66

TABLE 3 Summary of Uplift Mechanisms

Mechanism	Description Remarks, and Possible Tectonic Setting	Author or Proponent
1. Thermal expansion due to deep mantle plume or hot spot	Moving lithosphere overriding a hot spot or plume is heated: expansion causes uplift. Lithosphere may thin as well, and volcanism may follow melting. Hawaii is a good example; eastern Australia may be another.	Jackson and Shaw Morgan Crough, PUC
2. Thermal expansion due to overriding and subduction of ridge	Subduction of a midocean ridge or hot spreading center would introduce abnormally hot rocks into the upper mantle; expansion resulting could cause uplift and extensive volcanism.	Lipman-Christiansen DeLong et al.
3. Thermal expansion due to shear heating along lithosphere-asthenosphere interface	Viscous heating of shear zone between lithosphere and asthenosphere can produce expansion and resulting uplift.	Melosh Jackson and Shaw
4. Expansion accompanying partial melting	The volume of fusion for basaltic magma is about 8%; partial melting of the upper mantle would produce expansion.	
5. Hydration reactions such as serpentinization	Volatiles, introduced into a cool upper mantle could produce serpentine, with about 10% volumetric expansion. Hess suggested this for the Colorado Plateau in 1954.	Hess
6. Introduction of volatiles due to deep-seated dehydration of hydrous minerals	Humite, amphibole carbonate, and phlogopite are observed in deep-seated nodules in kimberlite; dehydration of these minerals and upward transport of the volatiles would contribute to or cause expansion and possibly partial melting.	McGetchin-Silver
7. Expansion due to depletion of 'fertile' mantle in garnet and iron resulting from basalt genesis	Subcontinental and suboceanic lithospheres (upper mantles) have similar densities and seismic velocities but very different temperatures.	Jordan
8. Crustal thickening due to horizontal transfer of mass in the lower crust	Process described by Gilluly prior to wide acceptance of plate tectonics to explain apparent mass deficiency under the Basin-Range and coincident elevation of the Colorado Plateau; details of process unspecified.	Gilluly
9. Underplating or subduction at very shallow angle	Metamorphic assemblages, such as high-pressure low-temperature eclogites suggest that ophiolite-like plates may exist beneath the Colorado Plateau; some believe that thrust angles could be nearly horizontal.	Heimstaedt-Schulze
10. Metamorphic (solid state) reactions as eclogite-basalt or spinel-olivine	Deep-seated solid-state reactions; would require heating, in general, to drive reactions to expansions.	O'Connell and Wasserburg Lovering Green-Ringwood
11. 'Simple' subduction and/or continental collision during subduction	High plateaus of South America and Tibet are examples of simple subduction processes. The Colorado Plateau may have been underlain by a subduction zone prior to 30 m.y., when the East Pacific Rise intersected the North American plate (Atwater, 1970).	Lipman-Christiansen Coney Damon, PUC
12. Cessation of subduction and resulting thermal equilibration of static slab	When subduction stops, the cold downgoing slab heats up. The resulting thermal equilibration and metamorphic reactions (both retrograde and prograde) are complex but will produce expansion, metamorphism, and possibly volcanism.	Thompson-Zoback, PUC Silver-McGetchin, PUC
13. Isolation of Colorado Plateau by listric normal faulting in surrounding areas	The Colorado Plateau is part of a much broader Rocky Mountain uplift. West, south, and east of the Colorado Plateau, crustal attenuation is achieved by stretching that is accompained by listric normal faulting. In this fashion the Colorado Plateau becomes isolated and remains intact. The listric normal faults are preferentially located along preexisting earlier thrust faults.	Bally, PUC
14. Lithospheric delamination, i.e., detachment and foundering of large pieces of subcontinental lithosphere	Owing to cooling, lithosphere detaches from crust and sinks into warmer, viscous, and denser asthenosphere. Counterflow of asthenosphere warms crustal rocks above; result is progressive uplift and volcanism.	Bird, PUC

PUC denotes Plateau Uplift Conference, Flagstaff, 1978.

prior to about 24 m.y. ago the Colorado Plateau was topographically low, had low relief and internal drainage, and had not yet developed an integrated through-flowing river system (the Colorado) which exited to the sea. There was significant faulting (including Basin-Range episode) in the Miocene between 18 and 20 m.y. ago, following which time the drainage became integrated into the ancestral Colorado River system (Pliocene time), which flowed off the plateau to the west. The Peach Spring tuff is a key marker unit in pinning down these relations. It appears that prior to eruption of this ash flow the Basin-Range province stood topographically high relative to the Colorado Plateau; after about 10 m.y. the reverse was true. The bulk of the regional uplift, some 1 km, occurred in the Miocene during this 8-m.y. span (between about 18 and 10 m.y. ago), and the Colorado Plateau province has stood high since. The regional uplift is obscured by contemporaneous Basin-Range faulting and relative subsidence west of the plateau.

Lucchitta has pointed out that an important measure of the uplift of the plateau with respect to sea level is provided by upwarping and faulting of sea level or near sea level deposits from the mouth of the Colorado to the mouth of the Grand Canyon. He concludes (1) that before Basin-Range faulting the plateau was lower structurally and topographically than the country to the west; (2) that at the end of Basin-Range faulting and deposition (about 8 m.y. ago) the Plateau was 1.1 km higher than the top of the adjacent basin fill, 4.3–5.7 km higher than the floor of adjacent basins, but still lower than nearby ranges; and (3) that since about 5 m.y. ago, at least 880 m of uplift of the plateau and Basin-Range have occurred through upwarping and faulting.

There was a general consensus that the plateau we see in Arizona today has been inherited from the modification of a regional northeast-sloping surface related to a broader uplift of much of the western United States dating from the Laramide. This old surface, extensively eroded in early Tertiary time, is covered by Oligocene volcanics on the western and southern margin of the plateau (Peirce; Young).

Lake Mead Region

The premeeting field trip route (led by Ivo Lucchitta and Richard Young), in an east-west direction, crossed Lake Mead and turned southward along the southwestern margin of the Colorado Plateau from the mouth of the Grand Canyon to U.S. Interstate 40. In this area, styles of tectonic deformation vary from strike slip faulting in the north and west to simpler normal faulting in the south and east (Figure 3). Principal features include Laramide (65 m.y.) plutons, early to middle Tertiary uplift, erosion, volcanism, and faulting. The Peach Springs tuff is a distinctive widespread stratigraphic marker that crops out near the margin of the plateau. This unit and the Miocene and Pliocene Horse Spring Formation and Muddy Creek Formation are critical for understanding the post-middle Miocene structural history of the major plateau margin faults; the structural and stratigraphic relationships of these units thus were focal points of the trip and of the meeting because they define the time of uplift and position of the boundary of the Colorado Plateau relative to the Basin-Range province.

Conclusions

The conference closed with a lively discussion, led by Bert Bally, Kevin Burke, and E. M. Shoemaker, which focused on the question: What new data do we need to address crucial questions for the Colorado Plateau? There was a consensus that the answers apply to all plateau uplifts of the world and that they will come from many disciplines. Important new data are likely to be contributed by (1) geologic and geomorphologic mapping of relations on the margins of uplifts, especially supported by absolute age determinations of associated volcanics which will constrain the amount and timing of uplifts; (2) petrology of volcanic rocks as clues to their origin and relationship to subducting plate margins; (3) petrology of xenoliths in volcanic rocks which provide important clues to the composition and state of the lower crust and upper mantle and, in particular, metamorphic events occurring at depth through quantitative constraints on pressure, temperature, and chronology of these rocks (there is a critical need for age-dating methods applicable to mafic and ultramafic lower crustal and upper mantle xenoliths); and (4) geophysics which will continue to provide the bulk of data for the physical properties of the lower crust and upper mantle.

New programs such as Continental Drilling, COCORP, satellite geodesy (e.g., NASA's proposed geodynamics program), and consortia approaches to topical studies may help considerably in providing new data.

Several conclusions were reached. Upper mantle structure, composition, and thermal processes are the key to understanding the ultimate causes of uplift in virtually every case. It was generally agreed that the Colorado Plateau could not be separated from the regional uplift of western North America during Tertiary time. The uplift of the Colorado Plateau province and adjacent regions can be bracketed in time between about 10 and 18 m.y., and the magnitude of upward movement was about 2 km. Available data (*P*, *S* velocity structure, *Q* electrical resistivity profiles, heat flow, and petrology of volcanics and xenoliths) strongly suggest partial melting within the upper mantle under much of the western United States. The 'Colorado Plateau uplift problem' then becomes one of explaining (1) the ultimate cause of the regional uplift of the American West and (2) why the Colorado Plateau survived as a structural and topographic entity while the country surrounding it was being severely modified by faulting and extensive erosion. It is apparent that the structural and topographic unit known as the Colorado Plateau is being 'nibbled' away from all sides by active erosion and faulting.

It was pointed out, partly in jest, by George Thompson that Dutton in 1892 suggested that uplifts are due primarily to expansion. No less than 14 fundamental uplift mechanisms were suggested (see Table 3), essentially all involving subduction, plate motion, and/or expansion due to heating or partial melting. The extensive volcanism in and around the Colorado Plateau is clear evidence that the upper mantle was being heated by some process; geophysical data on the current state of the upper mantle support this view. The wealth of data emerging from many disciplines is constraining the many possibilities for models of uplift mechanics. This conference was a dialogue among specialists of widely different disciplines largely focused on the Colorado Plateau; at the conclusion it was clear that we have

made substantial progress in the last decade but there was a clear consensus on the need for new data. Comparisons to plateaus elsewhere on continents, on the sea floor, and even on other planets (the Tharsis uplift on Mars) provided an overview that uplifts are most commonly associated either (1) with subduction or its direct effects or (2) with deep-seated thermal disturbances which result in expansion and uplift.

Like the Colorado Plateau itself, the problem of plateau uplifts is being nibbled at from all sides—progress is steady, if not spectacular.

Acknowledgments

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The abstract volume with field guide is available on microfiche. Order from American Geophysical Union, 1909 K Street, N.W., Washington, D. C. 20006. Document E79-003: $1.00. Payment must accompany order. Copies ($1.00, $6.00 overseas) are also available from C. Watkins, Lunar and Planetary Institute, 3303 NASA Road One, Houston, Texas 77058. The proceedings of the conference will appear in *Tectonophysics* in 1979.

Enclosure 5

72

TWO PROBLEMS OF INTRA-CONTINENTAL TECTONICS: RE-ELEVATION OF OLD MOUNTAIN BELTS AND SUBSIDENCE OF INTRA-CONTINENTAL BASINS

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ABSTRACT

Two widespread features of the continental interiors of plates, re-elevated ancient mountain chains and intra-continental basins, imply considerable vertical displacement a long way from plate margins. The rates at which uplift and subsidence occur in these environments are sufficiently slow that seismic risk in them is generally low. Basin subsidence may occur for a variety of reasons but a coincidence in timing with plate-margin collisional episodes indicates the possibility that stresses transmitted from the plate-margin help in initiating subsidence. Although many intra-continental basins overlie rifts, cooling in response to the thermal perturbation represented by rifting is not generally the cause of basin subsidence.

INTRODUCTION

Studies by members of the Department of Geological Sciences at the State University of New York at Albany concentrate on the tectonics of the past. We have adopted the position that with the recognition of the plate structure of the lithosphere (Wilson, 1965) basic understanding of the way in which tectonic processes operate today has been achieved. We have examined ancient rocks and structures with a view to seeing how the processes that operated in the past resembled those active now and to assess whether observations on the ancient systems can help to illuminate the way in which the modern processes are happening. The contrast with the methods of seismologists is great, but the limitations that are imposed both by the short periods over which earthquakes occur and by the short length of the seismic record are well known (Allen, 1975). Our approach has, perhaps, been to some extent complementary.

Much of our effort has been spent in studying ophiolites, and this has proved fruitful in furthering understanding of processes at mid-ocean ridges and oceanic transform faults (see, for example, Dewey and Kidd, 1976 and Dewey and Karson, 1978). Here I speculate briefly on two general problems of intra-continental tectonics that emerge from studies of ancient rocks and that have some relevance to a conference on intra-continental seismicity because both types of phenomenon imply substantial displacements within continents at a long distance from active plate boundaries.

LOCALIZED RE-ELEVATION OF OLD MOUNTAIN BELTS

High mountain belts within continents today such as the Himalaya, Tibet, the Kun-Lun and the Tien-Shan are products of continental collision (Molnar and Tapponnier, 1975).

The history of elevation of these mountains since the collision between Asia and India about 40 m.y. ago has not, as yet, been worked out in detail but it is certainly complex. Evidence of antecedent drainage from Tibet to India shows that the Himalayas have gone up more recently than the Tibetan plateau and folding of Recent sediments in the Tsaidam basin (Şengör and Kidd, 1979) may indicate progressively younger, more northeasterly mountain building.

Global assessments of current relative plate motions (for example, Minster and Jordan, 1978) reveal that India and Asia continue to converge at a rate of about 60 mm/yr. The general elevation of the collisional mountains can be attributed to active tectonics resulting from this convergence. Erosion rates among the great collisional mountains of Asia are extremely high, and it has long been recognized that such erosion is rapid enough to remove mountain belts in a few tens of millions of years (see for a discussion of erosion rates, Blatt, et al., 1980, pp. 22-30). If the process of plate convergence responsible for the elevation of the collisional mountains of Asia were to cease erosion could remove all vestiges of the mountains within 100 million years.

An anomaly is that some quite old intra-continental mountain belts are elevated today. For example, the Appalachians of North America where the first convergent event happened about 450 m.y. ago and the last continental collision about 290 m.y. ago contain extensive areas over 1 km above sea level and reach, in two areas, an elevation of about 1.5 km. The trend of the old fold belt shows up clearly in topographic maps although some areas with the youngest tectonism - for example, the Piedmont province - are not the most elevated. In the Ural Mountains, where collision occurred about 290 m.y. ago, elevations in two areas also reach 1.5 km and a large areas exceeds 1 km above sea level.

The Caledonides of western Europe, where convergence ended as long ago as 350 m.y., also contain substantial areas higher than 1 km above sea level and reach a maximum elevation of about 1.5 km.

Two possible explanations of these elevated ancient mountains are (1) that estimated erosion rates are misleadingly high and (2) that the old mountain belts, having had their elevations rapidly removed by erosion, are re-elevated in response to renewed tectonic stress. Holmes (1965) concluded that the first hypothesis was valid and that erosion rates were particularly high at present because of the activities of man so that misleadingly high estimates of erosion rates are obtained. Workers in southeastern Europe and Asia Minor have been particularly impressed with the contribution of goats to rapid erosion (James Jackson, personal communication, 1979).

New light has been shed on the second hypothesis with the increasingly widespread observation that intraplate stresses are commonly compressional

(see, for example, papers at this conference and Sykes, 1976). This suggests the possibility that the old mountain belts serve to localize newly applied intra-plate stresses and owe their elevation to the development of compressional structures, thrust-faults at the level of brittle fracture, in response to the concentration of intra-plate stress.

It is intuitively easy to envisage that the rocks of mountain belts, even after the initial elevations have been eroded away, may contain weak zones along which strain can be concentrated when intra-plate stress is later applied. The oceanic parts of plate interiors being made largely of olivine, are much stronger than the continental parts and therefore less likely to fail in response to intraplate stress. The interiors of the continental parts of plates can be considered as broadly made up of Phanerozoic mountain belts and material representing older Precambrian, mountain belts.

A significant observation is that although considerable elevations do occur locally within the older terrains (for example, the Torngat Mountains of Labrador are made of gneisses yielding ages of up to 3.5 X 10^9 years and are 1.5 km high), Precambrian mountain belts are nowhere represented by generally elevated elongate chains like the Urals or the Appalachians. The oldest continuously elevated mountain belt is perhaps the Akwapim -Atacora chain stretching from Accra, in Ghana, to the Niger River. This belt formed by collision at the very beginning of the Phanerozoic about 600 m.y. ago (Burke and Dewey, 1972). The absence of elevated Precambrian mountain chains can be interpreted as indicating that although mountain belts may be reactivated during about the first 600 m.y. after their formation older mountain belts have taken up all the additional strain along their zones of weakness and are not susceptible to further reactivation. Sykes (1978) has pointed out that intraplate tectonics associated with the opening of new oceans are concentrated in the youngest orogenic belts on the ocean margins indicating that the older terrains are less susceptible to reactivation.

An implication of the foregoing discussion is that the elevation of old mountain belts is episodic. A mountain belt that remained elevated for 300 million years with active erosion would expose extremely deep crustal levels. The Appalachians and Caledonides typically expose rocks found no deeper than a maximum of 10-15 km. If the concept of episodic elevation is valid, it is important to be able to show that there are parts of Phanerozoic mountain belts that are not elevated and that this was true for other areas in the past. Much of the Hercynian of Europe forms low ground and parts of the Appalachians (in, for example, the Piedmont and Maritime Canada) are low-lying. There is also stratigraphic evidence that the Appalachians and Caledonides were not substantially elevated at times in the Mesozoic.

A further problem is: how do intra-plate stresses differ between the episodes of elevation of the old mountain belts and the times when they are not elevated? This presumably depends on the nature of the stresses acting within the plates. So little is known about these stresses that it is difficult to know where to begin. One consideration is that not only the magnitude of the stress but also its orientation may be important. Zones of weakness in old mountain belts are likely to run parallel to the trend of the belt and the response to compressional stresses applied parallel to

the belt may be more intense than the response to stresses applied perpendicular to it.

From the point of view of intracontinental seismicity it would appear that those currently elevated old mountain belts are likely to be only slightly seismically active. The amount of faulting needed to re-elevate a mountain belt lying at sea level across a width of 200 km to a central peak elevation of 1 km in 1 million years ignoring erosion is small. If 200 equal thrusts are used for the uplift the average vertical offset is only 10 m and the average repeat time for 10 cm of offset on each fault is 10,000 years so that one of the 200 faults should move 10 cm on the average every 50 years. Both the time of elevation and the number of faults involved are likely to be underestimated in this crude calculation.

SUBSIDENCE IN INTRACONTINENTAL BASINS

A second problem relates to the subsidence of intracontinental basins. These basins, lying within continents and remote from active mountain-belts, contain thicknesses of a kilometer or more of sediments.

Basins such as the Michigan, Paris, North Sea and Chad basins have been the subject of much recent theoretical discussion. Papers in Bott (1976) have been followed by detailed local studies, such as those on the Michigan Basin (Sleep and Sloss, 1978; Stakes, 1978; McCallister, et al., 1978; VanSchmus, 1978; Fowler and Kuenzi, 1978; Haimson, 1978) and theoretical analyses (e.g., McKenzie, 1978a). Interpretations of basin evolution as the result of lithospheric rupture and stretching (McKenzie, 1978a, 1978b), loading and cooling (Haxby, et al., 1976) are popular. Major problems emerge when these interpretations are applied to basins in which the sequence of evolutionary events is established by geological evidence. For example, in the North Sea rifting extended through several discrete episodes between early Permian and earliest Cretaceous times (~140 m.y. time span) and the development of the overlying broad "epeirogenic" basin has occurred during the Tertiary (Ziegler, 1978). No model, based on phenomena coextensive with the basin can account for all the properties of such an area and involvement of the Alpine collision about 1000 km away may prove important.

The idea that basins commonly overlie rifts (fig. 1) is being increasingly widely recognized (see, for example, Burke, 1976) and although in some cases basin subsidence follows as an immediate and therefore possibly as a thermal consequence of the rifting in many cases, subsidence does not take place until very much later. If this late subsidence is to be treated as a consequence of cooling then it becomes necessary to postulate a second thermal event. Unless there is some independent evidence of such an event the thermally controlled subsidence hypothesis is not attractive as an explanation.

An alternative possibility is that subsidence in some basins may be related to events at relatively remote plate margins. For example, the beginning of subsidence in the Michigan Basin (rifted 10^9 years ago) was contemporary with the collision of the Taconic island arc with the North American continent (in Trentonian time 450 m.y. ago, Church and Stevens,

BASIN	RIFTING Years Ago X 10^6	BASIN DEVELOPMENT Yrs. Ago X 10^6	KEY REFERENCE
CHAD	135 - 100	25 →	BURKE 1976
PARIS	~280	75 - 50	DEBELMAS (ed.) Geol. de la France 1976
MICHIGAN	~1,100	650? → 350 450 → 350	HAXBY *et al.*, 1976
S. OKLAHOMA	~650	310	HOFFMAN, *et al.*, 1974
NORTH SEA	~280 230 - 180 150 130	65 →	ZIEGLER Geol. en Mijn. 1978
WEST SIBERIA	230 ?	~160 - 70	RUDKEVITCH, *et al.*, 1976

Table 1. Basins overlie rifts but there is no obvious relationship in timing. Some basins begin to subside long after rifting and their subsidence cannot be controlled by the rift related thermal event.

1971) recorded in the Appalachian Mountain belt more than 1000 km away. Although limited Jurassic subsidence is recorded in the West Siberian basin following Triassic rifting (Rudkevitch, et al., 1976) the rate of subsidence increased in the Early Cretaceous following collision of the Okhotsk and Kolyma blocks with the Siberian platform some 2000 km farther east (Herron, et al., 1974).

Transmission of compressional stress through the continental lithosphere may therefore play a role in the development of intracontinental basins as it appears to in the re-elevation of ancient mountain belts. Subsidence rates in intracontinental basins averaged over their tens of millions of years of life are usually much less than uplift rates for re-elevated mountain belts and seismic risk is probably even lower in these areas.

BIBLIOGRAPHY

Allen, C.R., 1975. Geological criteria for evaluating seismicity: Geol. Soc. Amer. Bull., v. 86, p. 1041-1057.

Blatt, H., Middleton, G. and Murray, R., 1979. Origin of Sedimentary Rocks, Prentice-Hall, 672 pp.

Bott, M.H.P., 1976. Sedimentary basins of continental margins and craton. Developments in Geotectonics 12, Elsevier, Amsterdam, 314 pp.

Burke, Kevin, 1976. The Chad Basin: an active intra-continental basin. Tectonophysics 36, 197-206.

Burke, Kevin and Dewey, John F., 1972. Orogeny in Africa. In: Dessauvagie, T.F.J. and Whiteman, A.J. (editors) African Geology. Ibadan, 1970. Geol. Dept. Univ. Ibadan Nigeria: 583-608.

Church, W.R. and Stevens, R.K., 1971. Early Paleozoic ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences: Jour. Geophys. Research. v. 76, p. 1212-1222.

Dewey, J.F. and Kidd, W.S.F., 1977. Geometry of plate accretion. Geol. Soc. Amer. Bull., 88, 960-968.

Fowler, J.H. and Kuenzi, W.D., 1978. Keweenawan turbidites in Michigan (deep borehole red beds), a foundered basin sequence developed during evolution of a proto-oceanic rift system: Jour. Geophys. Res., v. 83, p. 5833-5843.

Haimson, B.C., 1978. Crustal stress in the Michigan Basin: Jour. Geophys. Res., v. 83, p. 5857-5863.

Herron, E.M., Dewey, J.F. and Pitman, W.C., 1974. Plate tectonics model for the evolution of the Arctic. Geology 2: 377-380.

Holmes, A., 1965. Principles of Physical Geology, second edition, 1288 pp. Nelson, Edinburgh.

Karson, J. and Dewey, J.F., 1978. Coastal Complex, western Newfoundland: an early Ordovician oceanic fracture zone: Geol. Soc. America Bull., v. 89, p. 1037-1049.

McCallister, R.H., Boctor, N.Z. and Hintze, W.J., 1978. Petrology of the spilitic rocks from the Michigan Basin deep drill hole: Jour. Geophys. Res., v. 83, p. 5825-5831.

McKenzie, D., 1978a. Some remarks on the development of sedimentary basins: Earth Planet. Sci. Lett., v. 40, p. 25-32.

McKenzie, D.P., 1978b. Active tectonics of the Alpine-Himalayan belt: the Aegean Sea and surrounding regions. Geophys. J. R. Astron. Soc., v. 55, p. 217-254.

Molnar, P. and Tapponnier, P., Cenozoic tectonics of Asia: effects of a continental collision, Science, 189, 419-426.

Rudkevich, M.Y., 1976. The history and dynamics of the development of the west Siberian platform. Tectonophysics 36, 275-287.

Şengör, A.M.C. and Kidd, W.S.F., The post-collisional tectonics of the Turkish-Iranian Plateau and a comparison with Tibet. Tectonophysics 55, 361-376.

Sleep, N.H. and Sloss, L.L., 1978. A deep borehole in the Michigan Basin: Jour. Geophys. Res., v. 83, p. 5815-5819.

Stakes, D.S., 1978. Mineralogy and geochemistry of the Michigan deep hole metagabbro compared to seawater hydrothermal alteration: Jour. Geophys. Res., v. 83, p. 5820-5824.

Sykes, L.R., 1978. Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism and other tectonism post-dating continental fragmentation: Rev. Geophys. Space Phys., v. 16, p. 621-688.

VanSchmus, W.R., 1978. Rb-Sr geochronologic analysis of metagabbro at the bottom of the Michigan Basin deep drill hole. Jour. Geophys. Res., v. 83, p. 5832.

Wilson, J.T., 1965. A new class of faults and their bearing on continental drift. Nature, v. 207, p. 343-347.

Ziegler, P.A., 1978. Northwestern Europe: tectonics and basin development. Geologie en Mijnbouw, v. 47, p. 584-626.

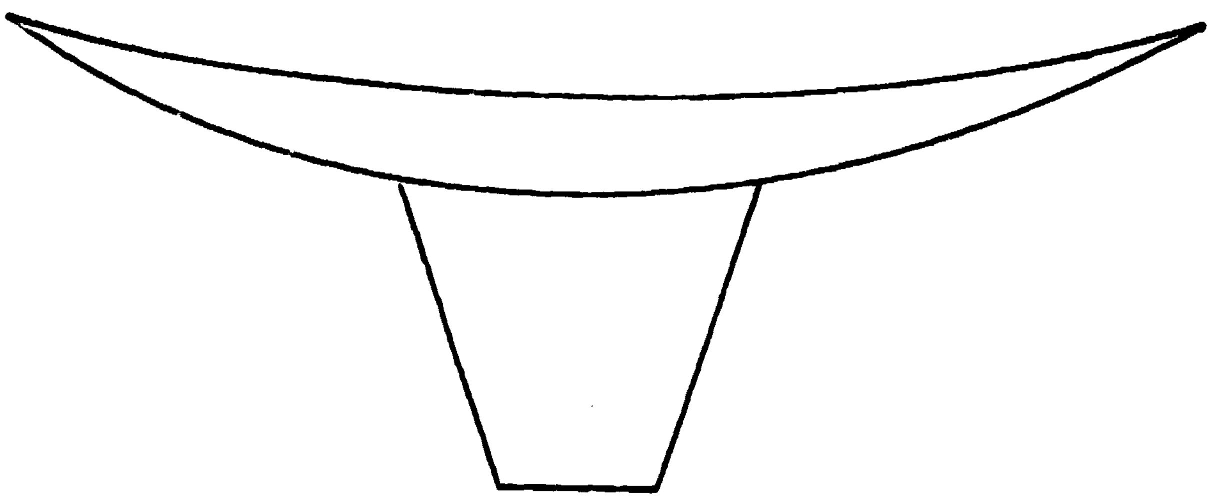

Figure 1. Sketch illustrating how intracontinental basins commonly overlie rifts. The resemblance to a bovine head has led to this being known as the 'long-horn condition' in Texas.